(1)

DARPA-NMR-81-01

AD A114654

A TECHNICAL ASSESSMENT OF SEISMIC YIELD ESTIMATION

APPENDIX

PART 2 OF 2

APPROVED FOR PUBLIC RELEASE
DISTRIBUTION UNLIMITED

JANUARY 1981

DEFENSE ADVANCED RESEARCH PROJECTS AGENCY
DARPA

1400 WILSON BOULEVARD
ARLINGTON, VIRGINIA 22209

RONDOUT ASSOCIATES, INCORPORATED

P.O. Box 224, Stone Ridge, New York 12484

Paul W. Pomeroy

January 10, 1980

Defense Advanced Research Projects Agency
1400 Wilson Boulevard
Arlington, Virginia 22209
Att'n: Col. George Bulin

Dear Col. Bulin;

In accordance with Dr. Romney's letter of 18 December 1979, I am submitting herewith one (1) copy of my summary statement on "Definition of Surface Wave Magnitude". In the course of reviewing the background material for this statement and the related issue of seismic yield estimates from surface wave magnitudes, a clear need for additional studies in the following areas became apparent.

1. Surface Wave Magnitudes. A systematic reevaluation of surface wave magnitudes should be undertaken if these magnitudes are to be used in the future:

a. Formulations using Love wave data should be derived and evaluated.

b. For M_{SR} and M_{SL}, surface wave amplitude envelopes and resultant path corrections for specific paths should be evaluated for source regions and stations of greatest interest.

c. M_{SR} and M_{SL} formations clearly will differ from region to region and the m_R term in the magnitude formulation of Båth should be evaluated.

d. With redefined and recalibrated M_{SR} and M_{SL} values, a reevaluation of M_S vs. Y should be carried out.

I feel strongly that with improved capability of assigning M_S, the uncertainty in yield can be significantly reduced.

2. Spectral Methods of Evaluating Long Period Energy Content. The current state of the art in surface wave magnitude determination is appalling. With the advent of digital data recording and handling, amplitude spectra for Love and Rayleigh waves should be available for a large number of events in the areas of interest. After correction for dispersion effects, these spectra should provide a major improvement in our ability to determine relative size of seismic events and ultimately lead to improvements in seismic yield estimates.

27

3. Instruments and Regional Data. One of the major sources of confusion in the magnitude picture has resulted from the difference in instrumentation between various countries. For intermediate (8-14 seconds) and long periods (> 10 seconds), the differences between the Kirnos instrumentation in the USSR and the narrow band long period instruments in this country are particularily striking. A major research effort should be initiated to study improvements in capability (defection, discrimination and yield determination) if broad band high gain instruments with significant response in the 8-14 second period range (part of the NSS bands) are utilized. Using digital data, this 'window' between the 4-9 second microseisms and the 14-22 second microseisms could be adjusted for maximum band width in a time dependent manner. In additional to potential improvements in detection of fundamental mode surface waves, investigations of the relative generation of higher mode Love and Rayleigh waves (predominantly in this period range) by explosions and earthquakes could enhance discrimination capability. Differences in the coefficients of the distance dependent terms in the magnitude formulations of Basham and Evernden are probably related to instrumental differences. The resolution of these differences will result in refined M_{SR} and M_{SL} measurements and thus in yield measurements.

Parenthetically, I might add that my review of the open literature also convinces me that, in spite of the formidable obstacles of 'tectonic contamination' and near source effects, the estimate of yield from surface wave measurements (either M_{SR}, M_{SL} or, more probably, by spectral methods) probably will provide more accurate estimates than body wave measurements.

In any case, I hope that the above comments are useful to you. Perhaps some of the suggested work is already done or underway. If so, I hope that unclassified aspects can be made available to the scientific community at large. I certainly enjoyed preparing this summary statement.

Sincerely yours,
RONDOUT ASSOCIATES, INCORPORATED

Paul W. Pomeroy
President

cc. Ralph Alewine, DARPA
William J. Best, AFOSR

Definition of Surface Wave Magnitude

Paul W. Pomeroy

11 January 1980

Rondout Associates, Inc.
P.O. Box 224
Stone Ridge, New York 12484

Tel. 914-687-9150

Definition of Surface Wave Magnitude

Introduction.

In the context of nuclear explosion seismology, surface wave magnitudes, M_S, derived from measurements of Love or Rayleigh waves, have assumed particular importance since they are utilized in two areas, namely:

(1) Discrimination - where the discrimination between explosions and earthquakes relies heavily on the contrast in surface wave generation by the two event types. This contrast is normally measured by M_S vs. body wave magnitude, m_b, criteria.

(2) Determination of explosion yield using M_S.

In this paper, the background of surface wave magnitude determination is outlined, followed by discussions of present practice in surface wave measurements at both teleseismic and regional distances. A final section outlines recommended studies related to improving determination of surface wave magnitudes and other, more sophisticated, methods of evaluating long period energy content in the seismic surface wave train.

Historic Background.

Following the successful development of the local magnitude (M_L) scale (Richter, 1935), magnitudes for California earthquakes, as recorded at Pasadena, were determined on a relatively routine basis. The M_L scale was originally defined for the California area only and for epicentral distances up to 600 km. In order to assign magnitudes at teleseismic distances, Gutenberg and Richter (1936) and Gutenberg (1945) developed a magnitude scale for shallow focus events at teleseismic distances "using the computed horizontal ground amplitude, in microns, of the large surface waves with periods near 20 seconds which form the maximum of the normal seismogram (the two horizontal components are combined vectorially)" Richter (1958). This surface wave formulation was in the form:

$$M_S^G = \log A_H - \log B + C + D \qquad (1)$$

where A_H is the vector combination of the horizontal ground motion,

B is the same quantity for a shock of magnitude zero and depends only on distance for a given focal depth,

C is the station correction factor and,

D depends on the depth of focus, the source radiation pattern, the absorption along the propagation path and crustal structure, together with any irregularities along the path.

This formula was developed empirically from recordings of 52 seismic events (all but five from the Circum Pacific Belt) as recorded at stations in Southern California and was based on the common (but not universal) observation that the maximum amplitudes of surface waves were associated with periods around twenty (20) seconds. Because amplitudes were measured on horizontal instruments, measurements of either Love or Rayleigh waves or both could be involved in this scale. This formula was considered applicable at epicentral distances between 4 and 180°. The term - logB as tabulated by Gutenberg is presented in Figure A-1 of the Appendix to this report. Revised values published by Richter, 1958, are shown in Figure A-2. Since little information on C or D was (and is) available, magnitudes were routinely determined from the first two terms of Eq. 1. This formula is sometimes written as $M_S G = \log_{10} A + B(\Delta)$ (2). We will use this B (Δ) terminology later.

In 1952, Båth (1952) investigated the use of the vectorial component of surface waves (i.e. only Rayleigh waves) as a surface wave magnitude measurement tool. Using data from 305 earthquakes (141 of which were of shallow depth) which occurred throughout the world and were recorded at Pasadena, he derived

$$M_{SR}^{Båth} = \log A_Z - 1.66 \log\Delta + 0.0082h + m_R + C\,(M_o - M_{calc}) \quad (3)$$

where the first term accounts for the vertical ground amplitude, -logB is the distance dependent term and is equivalent to the Gutenberg value, the third term accounts for the focal depth (h), the fourth is a regional variation correction and the fifth accounts for the observation that the ratio of vertical to horizontal amplitudes apparently increased with the magnitude. This formula was developed for, and restricted to, wave periods between 17-23 seconds and was designed to agree with the M_S^G values derived from the same data set. It was reported

valid for focal depth (h) $\leq$ 100 kms. Because of a lack of data on the last three terms, this formula was, and is, utilized using only the first two terms.

Therefore, at this point in time, it was realized and, to a varying extent, accounted for, that the following factors affected surface wave magnitude determination:

1. The depth of the source (D in Gutenberg's formula and .0082h in Båth
2. Station corrections (for structure near the station and instrument effects - C in Gutenberg's formula).
3. Surface wave radiation patterns and source effects (D in Gutenberg's formulation.)
4. Regional variations in propagation resulting from differences in absorption and wave path irregularities (m_R in Båth and D in Gutenberg).

In practice, however, magnitudes were assigned using various formulations, and it was often impossible to compare magnitude values obtained for the same earthquakes.

Current Surface Wave Magnitudes.

An important milestone in the development of surface wave magnitude scales was published by Vanek et al. (1962) who proposed the following formula:

$$M_S^P = \log (A/T)_{max} + \sigma(\Delta) \qquad (4)$$

where $\sigma(\Delta)$ the calibrating function was assumed to be of the form a log Δ + b with a = 1.66, b = 2.0. a and b were weighted averages from 14 individual calibrating functions suggested in the literature. Since the calibrating functions were all for periods around 20 seconds, log 20 (or 1.3) was added to the expression. The final formulation was:

$$M_S^P = \log (A/T)_{max} + 1.66 \log \Delta + 3.3 \qquad (5)$$

This was referred to as the Prague formula and was deemed useful between 2° and 160° epicentral distance. Technically, this formulation was for horizontal surface waves. It incorporated the period of ground motion, T, to account for the following:

1. In continental propagation, the observed period associated with maximum amplitudes is not 20 seconds but is controlled by the group velocity dispersion curve for the particular propagation path.

2. Using broad-band Kirnos type instruments, with the significant response in the 1-12 second range, the maximum amplitudes were recorded at periods different from 20 seconds as in 1.

Note that this formulation does not take into account some of the specific items relating to variations in surface wave amplitudes particularly depth which were left for later study but the formula was considered useful for determining the surface wave magnitudes in an uniform manner on a world wide basis.

IASPEI Definition.

In 1967, the Committee on Magnitudes of the International Association of Seismology and Physics of the Earths Interior (IASPEI) recommended the general adoption of a new surface wave formula (essentially the Vanek et.al. or Prague formulation). $M_S^P = \log (A/T) + 1.66 \log \Delta + 3.30$ (6)
with the A/T term referring to maximum amplitude and corresponding period of the vertical component Rayleigh wave (Båth, 1967). This formula was adopted by the United States (USCGS at that time), Canada and many other workers in the field (Basham, 1971).

The $1.66 \log \Delta + 2.0$ in this formulation (based on the average of 14 different calibrating functions) is in good agreement with the Gutenberg, emperically determined, $B(\Delta)$ term ($1.656 \log \Delta + 1.818$, Fig. A-1) and is in ever closer agreement with values of the $B(\Delta)$ term published by Richter (Fig. A-2 of this report). All of these relationships imply that the amplitude decrement with distance of 20 second waves due to geometric spreading, dispersion and adsorption is proportional to $\Delta^{-5/3}$ power. All recent M_S studies have shown that, when applied to 20 second waves recorded at epicentral distances greater than approximately 25°, this distance correction will yield M_S values that are essentially independent of distance. However, the M_S values will still be strongly dependent on source and propagation effects (Marshall and Basham, 1972).

In any case, this formula is the basic surface wave magnitude formula in use throughout the world today and is the one that, for distances greater than 25°, should be used as a starting point for refinement of surface wave scales for application in specific source regions.

Surface Wave Magnitudes at Distances less than 25°.

Because of the importance of surface wave magnitudes in $M_S:m_b$ discrimination criteria, numerous additional investigations of the surface wave magnitudes were initiated. Basham (1969, a and b) reported that a great improvement in identification using small Rayleigh waves is achieved if the recording stations and the source are seperated by purely continental paths and the measurements for surface wave magnitudes are restricted to the shorter period (8-14 second) Rayleigh waves commonly identified as Rg. Basham (1971), however, reported that the M_S values derived using these waves differed significantly from those computed using the Gutenberg formulas. Both he and Evernden (1971) examined the validity of the distance calibration function in Eq 6 as applied to short period (8-14 seconds) waves. Evernden used data from earthquakes and explosions in the United States as recorded on LRSM long period narrow band instruments while Basham used a number of Canadian stations with wider band instrumentation approximately equivalent to that of WWSSN stations. Basham's measurements were in the period range 8-14 seconds while Evernden's data were at 10-14 second periods for explosions and 17-19 seconds for earthquake Rayleigh waves. Both authors noticed a systematic dependence of M_S on distance implying that the relationship with distance in Eq. 6 was not appropriate at these smaller ($< 25°$) distances in North America. Basham suggested that the appropriate coeffecient in the log $\underline{\Delta}$ term was .79 while Evernden suggested approximately 1.0. Basham (1971) proposed the formula

$$M_S^{B1} = \log (A/T) + 0.79 \log \underline{\Delta} + 4.54 \qquad (7)$$

although the formulation was modified in the Marshall and Basham (1972) paper discussed below. Evernden et.al. (1971) proposed the formula

$$M_S^{E} = 0.92 + \log \underline{\Delta} + A/T \qquad (8)$$

A constant factor of 3.0 should be added to this formula to make it comparable to M_S^{B1}, M_S^{B+M}, M_S^{N+K}, etc. because of differences in amplitude units used. The important result is that, for the shorter period Rayleigh waves, at least for North American events observed in North America at distances $< 25°$, the coefficient of the log $\underline{\Delta}$ term in the magnitude equation is of the order 0.8-1.0, not 1.66.

Two things should be pointed out:

1. Different values of the coefficient should be used if differing instruments are used for the observations.
2. The coefficients may be different on different continents.

Marshall and Basham (1972) have conducted a detailed reanalysis of Rayleigh wave surface wave magnitudes and the components that enter into the evaluation of a surface wave magnitude. That study is outlined in paragraphs 1-3 below.

1. The term involving log Δ. The coefficient of 1.66 of the log Δ term in M_S^P indicates that the amplitude decrement with respect to 20 second waves due to geometric spreading, dispersion and absorption is proportional to $\Delta^{-5/3}$. Typical continental Rayleigh wave group velocity curves, particularly in North America show a minimum as a period of approximately 12 seconds. For epicentral distances up to 25°, if frequency dependent absorption and scattering have not affected the periods around 12 seconds, the maximum amplitude in the Rayleigh wave train observed on wide band long period seismographs measured to compute M_S will be an Airy phase at 12 seconds. If so, the dispersion effects are negligible and the geometric spreading term is proportional to $\Delta^{-5/6}$. For reasonable values of Q, absorption is negligible and the appropriate distance correction term in M_S is .8 log Δ. The authors propose, then, that for distances up to 25° the distance dependence term $B'(\Delta)$ be proportional to 0.8 log Δ ($B'(\Delta)$ is tabulated in A-3 of the appendix) and at large distances be the same as Gutenbergs (Fig. A-1). The absolute level of this term is adjusted so that magnitude determinations from the formula will give results equivalent to those for M_S^P derived from seismograms at large epicentral distances ($\geq$ 25°). The values of this term are appropriate to displacements defined in nm. $B'(\Delta)$ will be about 3.0 smaller than the distance correction for M_S^P based on displacement in microns.

2. Wave period T. The period of the waves with maximum ground amplitude depends on the spectral characteristics of the source, the dispersion properties of the transmission path and the absorptive properties of the path. The overriding effect is related to the dispersion. For a mean of 30 group velocity curves in Eurasia, 4 in North America and 1 each for a mixed continental oceanic path and an oceanic path, the authors derived amplitude envelopes for vertical component Rayleigh waves (Figure A-4 of the Appendix) and path corrections based on these data (P(T) in Appendix , Figure A-5).

3. <u>Depth of Source</u>. As the focus of a seismic event gets deeper, the Rayleigh wave train generated and observed on the surface at a distance becomes progressively less rich in high frequency energy, i.e. the observed spectrum peaks at a progressively longer period. Coupled with this effect will be the spectral characteristics of the source itself; and the two effects cannot be easily seperated. For explosions, the source depth only varies by a few kilometers and the source depth effects will be minimal. For earthquakes, the effect should be measureable. A technique using NOS depths for calibration was devised and used in this study. The depth correction is then assumed to be 0.008h from Båth (1952) which normalizes M_S to the equivalent value for a surface focus event. With a good depth data set, this correction might be useful but, at present, it is of limited significance.

Finally then, Marshall and Basham derive a Rayleigh surface wave magnitude.

$M_{SR}^{M+B} = \log A + B'\Delta + P(T)$ (9) from 1 station and a final

$M_{SR}^{M+B} = M_S + 0.008h$ (10)

where $B'(\Delta) + P(T)$ are given in Tables A-1 and A-3.

Eliminating the depth correction, this M_{SR}^{M+B} is perhaps the best formula available for use in North America and Eurasia at distances less than 25°, if measurements are made in the 8-14 second period range. It should serve as a starting point for derivations of M_S for particular regions as recorded at specific stations.

Nuttli (19¯¯) showed that the 1.66 log Δ representation in M_S^P represents a linear approximation to a theoretical attenuation curve (Ewing, et.al. 1957) for 20 second period surface wave amplitudes in the time domain

$$A = A_0 \Delta^{-1/3} \sin \Delta^{-1/2} \exp(-0.015\,\Delta) \qquad (11)$$

where the value of 0.015 deg^{-1} satisfies the original empirical data of Gutenberg and Richter (1936). Nuttli (1973) made a set of measurements of surface waves with periods of 3-12 seconds in eastern North America and indicated that these measurements in the distance range 2-20° could be adequately represented by a straight line with slope -1.66. At first glance, this appears to contradict the work of Basham (1971) and Evernden (1971). However, Nuttli's observations are

not made at the Airy phase; rather he makes several measurements of amplitude and period throughout the vertical component wave train and picks the maximum value of (A/T). Basham and Evernden apparently measure the maximum amplitude observable on the seismogram and the period associated with that maximum. Thus, Nuttli is, in general, treating a non-Airy phase and measuring $(A/T)_{max}$ whereas the other authors are measuring an Airy phase and (A_{max}/T).

In a later paper, Nuttli and Kim (1975) show that the slope of -1.66 does not fit the theoretical curve of Eq. 11 at distances less than 25°. The slope is more appropriately -1.07 and the appropriate equation for M_S in the distance range 10-30°, <u>if M_S is to be determined from 20 second period waves</u>, is

$$M_S^{N+K} = 4.16 + 1.07 \log \Delta + \log A/T \qquad (12)$$

The value 4.16 was determined by requiring Eqs. 12 and 6 to give the same values of M_S at $25° \leq \Delta \leq 30°$.

Nuttli and Kim also considered that, <u>for earthquakes</u> at near to regional distances, the M_S values obtained from Rayleigh and Love waves using their formula, are identical. This point requires further investigation. Thus for M_S measurements of surface waves at distances less than 25°, <u>if measurements are made at periods around 20 seconds</u>, the Nuttli and Kim (1975) formula is preferable. It is perhaps preferable to make 20 second period measurements for at least three reasons.

1. No scaling to original magnitude formulas is required
2. 20 second waves do not 'feel' the effects of small changes in upper crustal structure to the same degree that Airy phase measurements do. Thus, $M_{S(20)}$ will provide more stable estimates of size or yield.
3. 20 second waves from most earthquakes can be observed on long period instruments with suitable response (i.e. such as the HGLP) down to extremely low magnitudes (see Evernden (1975)).

Conclusions.

1. At teleseismic distances greater than 25°, the IASPEI, Prague or Gutenberg formulas are the basis for calculating the surface wave magnitudes

$$M_{SR}^{P} = \log (A/T) \max + 1.66 \log \Delta + 3.30 \text{ (for } \Delta \geq 25°) \qquad (6)$$

using Rayleigh waves recorded on vertical component seismographs.

2. At distances less than 25°, when the source and receiver are on the same continental unit (valid only for North America and Eurasia) the Basham and Marshall formulation should serve as a starting point for measurements of M_S for Rayleigh waves.

$$M_{SR}^{M+B} = \log A + B' \Delta + P(T) \qquad \text{(for } \Delta \leq 25°) \qquad (9)$$

if measurements are made in the 8 - 14 second period range.

3. At distances less than 25°, if M_S measurements are made at periods around 20 seconds the formula should be

$$M_S^{N+K} + 4.16 + 1.07 \log \Delta + \log A/T \qquad (12)$$

4. With the possible exception of M_S^{P}, none of the above are applicable to Love waves without additional verification, although they are routinely used in M_{SL} determinations.

Recommendations.

If M_S (R or L or both) will continue to be useful both in the M_S:m_b explosion-earthquake discriminant analyses and in determination of explosive yield, several studies should be carried out.

1. A complete reevaluation of surface wave magnitude determination and their calibrations should be carried out. Specifically,

   a. Surface wave magnitudes should be derived on a region to region basis for each recording station of interest for events in the Eurasian continent following the technique outlined by Marshall and Basham, i.e., surface wave dispersion curves should be drived for each region to recording stations and the resultant P(T) term used for magnitude calculations. The above methodology should be used if measurements are made in the 8 - 14 second period range. Alternatively, if M_S measurements are to be made at periods around 20 seconds, the Nuttli and Kim formulation should be evaluated and utilized.

   b. Love waves should be utilized for independent surface wave magnitude determinations. For distances greater than 25°, the IASPEI formula can be used directly. For distances less than 25°, dispersion curves and amplitude envelopes for continental Love waves in the period range 8-60 seconds should be obtained and appropriate path corrections $P_L(T)$ determined. The applicability of the distance term 0.8 log Δ should be evaluated in order to determine M_{SL} on a region by region basis.

   c. Three component broad band intermediate period instruments should be developed, deployed and utilized for determination of surface wave magnitudes at distances less than 25° on the same continental land mass since the maximum A/T values will usually occur around 8-14 seconds in this distance range.

2. With the advent of broad band seismic systems recording digitally, routine coordinate rotation and independent determination of M_{SR} and M_{SL} become possible. More vitally, it becomes possible to consider separately the entire Rayleigh and Love amplitude spectra. For distances $<25°$ or $>25°$, using data for Love and Rayleigh wave dispersion along particular paths of interest, the path effect can be removed. The absolute level of the resultant spectra (Love and Rayleigh separately) can be used to rank the

yield or size of the event and the spectral method is far superior to measurements at one particular period. (Grant and Mansinha, 1977). The development of spectral methods (Love and Rayleigh) for determination of event size and their calibration for paths of interest should be undertaken and will result in major improvements in yield estimates and in discrimination.

References.

Basham, P.W. (1969a), "Canadian Magnitudes of Earthquakes and Nuclear Explosions in South-Western North America", Geophys. J.R. astr. Soc., 17, 1.

Basham, P.W. (1969b), "Canadian Detection and Discrimination Thresholds for Earthquakes and Underground Explosions in Asia", Can. J. Earth Science, 6, 1455.

Basham, P.W. (1971), " A New Magnitude Formula for Short-Period Continental Rayleigh Waves", Geophys. J.R. astr. Soc., 23, 255-260.

Båth, M. (1952), "Earthquake Magnitude Determination from the Vertical Component of Surface Waves", Trans. Am. Geophys. Un., 33, 81-90.

Båth, M. (1967), "Recommendations of the IASPEI Committee on Magnitudes", Seismological Bulletin, May 1967, Seismological Institute, Uppsala, Sweden.

Evernden, J.F. (1971), "Variation of Rayleigh Wave Amplitude with Distance", Bull. Seism. Soc. Am., 61, 231-240.

Evernden, J.F. (1975), "Further Studies on Seismic Discrimination", Bull. Seism. Soc. Am., 65, 359-392.

Evernden, J.F., W.J. Best, P.W. Pomeroy, T.V. McEvilly, J.M. Savino and L.R. Sykes (1971), "Discrimination Between Small-Magnitude Earthquakes and Explosions, J. Geophys. Res., 76, 8042-8055.

Grant, J.A. and L. Mansinha (1977), "Seismic Magnitude from Fourier Analysis", Bull. Seism. Soc. Am., 67, 453-462.

Richter, C.F. (1935), "An Instrumental Earthquake Scale", Bull. Seism. Soc. Am., 25, 1-32.

Gutenberg, B. (1945), "Amplitudes of Surface Waves and Magnitudes of Shallow Earthquakes", Bull. Seism. Soc. Am., 35, 3-12.

Gutenberg, B. and C.F. Richter (1936), "On Seismic Waves (third paper)", G. Beitr., 47, 73-131.

Marshall, P.D. and P.W. Basham (1972), "Discrimination Between Earthquakes and Underground Explosions Employing an Improved M_S Scale", Geophys. J.R. astr. Soc., 28, 431-458.

Nuttli, O.W. (1973), "Seismic Wave Attenuation and Magnitude Relations for Eastern North America", J. Geophys. Res., 78, 876-885.

References (Cont.)

Nuttli, O.W. and S.G. Kim (1975), "Surface Wave Magnitudes of Eurasian Earthquakes and Explosions", Bull. Seism. Soc. Am., 65, 693-709.

Richter, C.F. (1958), Elementary Seismology, W.H. Freeman, San Fransisco.

Vanek, J., A. Zatopek, V. Karnik, N.V. Kondorskaya, Y.V. Riznichenko, E.V. Savarensky, S.L. Solov'ev and N.V. Shebalin (1962), "Standardization of Magnitude Scales", Bull. (Invest.) Acad. Sci. USSR Geophys. Ser., 2, 108-111.

APPENDIX

Figure A-1

TABLE 4

Revised Values of $-\log B$ in Equation (1)

(Δ = epicentral distance in degrees)

Δ	0	1	2	3	4	5	6	7	8	9
20	3.97	4.01	4.04	4.07	4.10	4.13	4.16	4.19	4.21	4.24
30	4.26	4.29	4.31	4.33	4.35	4.38	4.40	4.41	4.43	4.45
40	4.47	4.49	4.50	4.52	4.54	4.56	4.57	4.59	4.60	4.62
50	4.63	4.65	4.66	4.67	4.69	4.70	4.71	4.73	4.74	4.75
60	4.76	4.77	4.79	4.80	4.81	4.82	4.83	4.84	4.85	4.86
70	4.87	4.88	4.89	4.90	4.91	4.92	4.93	4.94	4.95	4.96
80	4.97	4.98	4.99	5.00	5.00	5.01	5.02	5.03	5.04	5.04
90	5.05	5.06	5.07	5.08	5.09	5.09	5.10	5.11	5.12	5.12
100	5.13	5.14	5.14	5.15	5.16	5.17	5.17	5.18	5.19	5.19
110	5.20	5.21	5.21	5.22	5.22	5.23	5.24	5.24	5.25	5.25

Δ........	124	128	130	135	140	145	150	160	162	165
log B.....	5.28	5.29	5.30	5.32	5.33	5.34	5.35	5.35	5.34	5.33
Δ.........	170	172	173	174	175	176	177	178	179	180
log B.....	5.32	5.31	5.30	5.28	5.25	5.22	5.20	5.15	5.1	5.0

TABLE 5

Values of $-\log b$

Δ	4	4½	5	5½	6	6½	7	7½	8
log b.....	4.60	4.74	4.87	5.00	5.12	5.23	5.33	5.43	5.52
Δ	8½	9	9½	10	10½	11	11½	12	12½
log b.....	5.60	5.67	5.74	5.80	5.85	5.90	5.95	6.00	6.04
Δ........	13	13½	14	15	16	17	18	19	20
log b.....	6.08	6.12	6.15	6.22	6.28	6.33	6.38	6.43	6.47

From Gutenberg (1945).

Figure A-2

Table 22-3 Logarithms of the Maximum Combined Horizontal Ground Amplitude *A* (in microns) for Surface Waves with Periods of 20 Seconds Produced at the Given Distance by a Standard Shock Taken as Magnitude Zero. (Correlation with Table 22-1 for short distances imperfect and now under investigation)

Δ (degrees)	−log A	Δ (degrees)	−log A
20	4.0	90	5.05
25	4.1	100	5.1
30	4.3	110	5.2
40	4.5	120	5.3
45	4.6	140	5.3
50	4.6	160	5.35
60	4.8	170	5.3
70	4.9	180	5.0
80	5.0		

From Richter (1958).

Figure A-3

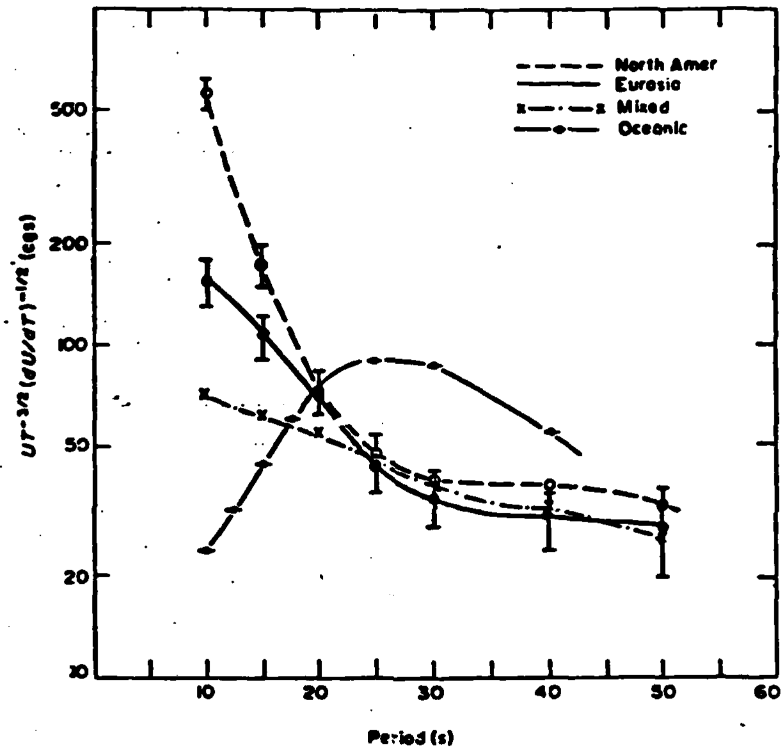


FIG. 2. Amplitude envelopes of vertical component Rayleigh waves as a function of period determined by the stationary phase approximation from the group velocity (U) versus period (T) curves for four transmission paths. The Eurasian and North American continental curves are the means of 30 and four determinations, respectively; standard deviation bars at each period are shown. The remaining two curves are examples for the particular transmission paths.

From Marshall and Basham (1972).

Figure A-4

Table 1

B′(Δ), A revised distance correction function for M_s

Δ°	B′	Δ°	B′	Δ°	B′	Δ°	B′	Δ°	B′
		20	1·25	40	1·57	60	1·85	80	2·08
1	0·17	21	1·27	41	1·59	61	1·87	81	2·09
2	0·35	22	1·29	42	1·61	62	1·89	82	2·10
3	0·57	23	1·31	43	1·62	63	1·90	83	2·11
4	0·67	24	1·32	44	1·64	64	1·91	84	2·12
5	0·78	25	1·34	45	1·65	65	1·92	85	2·13
6	0·84	26	1·36	46	1·66	66	1·93	86	2·14
7	0·90	27	1·38	47	1·68	67	1·94	87	2·15
8	0·95	28	1·40	48	1·70	68	1·95	88	2·16
9	0·98	29	1·41	49	1·71	69	1·96	89	2·17
10	1·02	30	1·43	50	1·72	70	1·97	90	2·18
11	1·05	31	1·44	51	1·74	71	1·98	91	2·18
12	1·08	32	1·45	52	1·75	72	1·99	92	2·19
13	1·11	33	1·47	53	1·76	73	2·01	93	2·20
14	1·13	34	1·48	54	1·77	74	2·02	94	2·21
15	1·15	35	1·50	55	1·78	75	2·03	95	2·22
16	1·17	36	1·52	56	1·80	76	2·04	96	2·23
17	1·19	37	1·54	57	1·82	77	2·05	97	2·24
18	1·21	38	1·55	58	1·83	78	2·06	98	2·25
19	1·23	39	1·56	59	1·84	79	2·07	99	2·26

Figure A-5

Table 2

Magnitude corrections for selected paths as a function of period, P(T)

		M_s Correction		
T (s)	Continental N. America	Continental Eurasia	Mixed Cont.–Oc.	Oceanic
10	−0·75	−0·30	+0·00	+0·50
11	−0·67	−0·27	+0·01	+0·45
12	−0·61	−0·24	+0·03	+0·38
13	−0·53	−0·21	+0·04	+0·33
14	−0·46	−0·18	+0·05	+0·27
15	−0·38	−0·15	+0·07	+0·20
16	−0·30	−0·13	+0·08	+0·15
17	−0·24	−0·10	+0·09	+0·09
18	−0·16	−0·07	+0·10	+0·03
19	−0·08	−0·04	+0·12	−0·03
20	+0·00	+0·00	+0·13	−0·09
21	+0·01	+0·03	+0·15	−0·08
22	+0·03	+0·05	+0·16	−0·07
23	+0·04	+0·07	+0·17	−0·06
24	+0·05	+0·11	+0·18	−0·05
25	+0·07	+0·14	+0·20	−0·04
26	+0·09	+0·18	+0·21	−0·03
27	+0·11	+0·22	+0·22	−0·03
28	+0·13	+0·24	+0·23	−0·02
29	+0·14	+0·27	+0·24	−0·01
30	+0·16	+0·30	+0·25	+0·00
31	+0·17	+0·32	+0·26	+0·01
32	+0·18	+0·33	+0·27	+0·02
33	+0·20	+0·34	+0·28	+0·03
34	+0·21	+0·35	+0·29	+0·04
35	+0·23	+0·36	+0·30	+0·05
36	+0·25	+0·37	+0·31	+0·06
37	+0·27	+0·38	+0·32	+0·07
38	+0·28	+0·39	+0·33	+0·08
39	+0·29	+0·40	+0·34	+0·09
40	+0·31	+0·41	+0·35	+0·10

Both from Marshall and Basham (1972).

University of California, San Diego
Keith Priestly

ESTIMATION OF SURFACE WAVE MAGNITUDE

Gutemberg (1945) defined surface wave magnitude as

$$M_s(G) = \log_{10}A + 1.66 \log_{10}\Delta + 2.0$$

where A is the amplitude of the resolved horizontal ground motion at a period of 20 seconds and $\log_{10}\Delta$ is a distance normalizing term which corrects for the effects of geometrical spreading, dispersion and absorption of the propagating surface waves. The formula was developed empirically for events recorded at epicentral distances greater than 20°, and was based on the supposition that the maximum trace amplitude corresponds to waves with periods of about 20 seconds. Period had no further significance in the determination of M_s. Vanck *et al.* (1962) proposed a revised definition for surface wave magnitude which considered those cases where the amplitude measurements were not made for 20 second surface waves. Their surface wave magnitude is given by the formula

$$M_s(P) = \log_{10}(\frac{A}{T}) + 1.66 \log_{10}\Delta + 3.3$$

where A is the maximum horizontal component Rayleigh wave with period T. When applied to surface waves with period near 20 seconds from the earthquakes at teleseismic distances, the the two methods yield results which agree to within 0.1 to 0.2 magnitude units.

When applied to short period surface wave data (8 to 14 second periods) commonly observed at regional distances, $M_s(P)$ is considerably longer than $M_s(G)$ (Basham, 1969, 1971; Evernden, 1971). This arises due to a greater propagation efficiency of shorter period waves over continental paths compared to mixed path types. Basham (1971) redefined surface wave magnitude for measurements made on short period continental

Rayleigh waves which is independent of distance. This is given by the formula

$$M_s(B) = \log_{10}(\frac{A}{T}) + 0.79 \log_{10}\Delta + 4.54$$

Evernden (1971), noting the same discrepancy in M_s determination, redefined M_s as

$$M_s(E) = \log_{10}(\frac{A}{T}) + B_s(\Delta)$$

where the distance correction term is given by

$$B_s(\Delta) = \begin{cases} 0.92 + \log(\Delta) & \Delta \leq 25° \\ 1.66 \log(\Delta) & \Delta \geq 25° \end{cases}$$

The period of the waves with maximum ground motion is dependent on the initial spectral distribution of energy at the seismic source, the source depth and the transmission characteristics of the path.

In a number of recent studies, Brune and his students have shown that the seismic moment, and hence the magnitude, is strongly dependent on period for moderate earthquakes. This results from the fact that the higher frequency body wave, hence m_b, are dominant by processes occurring during the initial rupture while the long period surface waves reflect the average rupture process.

The effects on surface wave amplitude of source depth involves the nature of the source, the fault orientation, and the earth structure in the vicinity of the source. As the source depth increases, the amplitude of short period surface waves decrease. In addition, holes in the spectrum, periods where surface wave energy cancels, can arise as a result of variation in source depth.

The dispersion characteristics of the transmission path have a considerable effect on shaping the amplitude envelope of the surface wave train. For example, if a group velocity minimum occurs for an

earth structure along a transmission path, the wave train can be undispersed over a range of periods resulting in a large amplitude Airy phase. Along another path whose earth structure contains a thick sedimentary section, the short periods are spread out in time leading to a dispersed wave train. Amplitude measurements direct from the seismogram would lead to a larger M_s value for the Airy phase than from the dispersed wave train.

The effects of surface wave attenuation are most variable at short periods. The current body of knowledge indicates 20 - 30% variation in Q for periods greater than 20 seconds, irrespective of path type resulting in .1 to .2 variation in M_s determined at those periods. For periods of 10 seconds, there can be as much as an order of magnitude variation in Q resulting in large variations in M_s.

AFTAC

SURFACE WAVE MAGNITUDE DEFINITIONS

Thomas D. Eisenhauer

Surface Wave Magnitudes. The major deficiency in current methodology for computing M_S is related to making amplitude measurements for LR. The variations in dispersion characteristics and multipath interference make it difficult to measure valid amplitudes near 20 seconds. The practice of correcting amplitudes at shorter periods to the amplitude at 20 seconds (Marshall and Basham, 1972), is better than no correction at all. However, Q for LR shows more variation over different paths for the shorter periods than for periods near 20 seconds (Mitchell, 1975) and dispersion and multipath effects are still a problem. Another problem with M_S is related to the relatively low S/N for LR compared to body waves. Some processing techniques can be used to good advantage to improve S/N for LR. I will comment on data, data processing, measurements distance normalization and corrections.

Data. The data should be in digital form. While both time and frequency domain measurements are desirable, spectral measurements offer distinct advantages for surface waves. Further, most data processing to improve S/N requires digital data. The network of stations should, when possible have continental paths to the test sites. The SRO and HGLP network along with long-period arrays (KSRS, ALPA and NORSAR) should provide the best data for LR measurements.

Data Processing. Frequency fi[illegible] along with phase matched filtering (Herrin, 1976), have potential for improving S/N. The purpose of this filtering is to produce a seismogram with high S/N and a close approximation of the fundamental mode Rayleigh waves traveling over the least time path from source to station.

Measurements. I recommend the use of spectral measurements for Rayleigh waves. Such measurement circumvent some of the problems with variations in dispersion, and simplfy the distance normalization term in the magnitude equation. I believe that the best spectral amplitudes can be obtained in the period range of 18 to 22 seconds since Q appears to be stable at these period (Mitchell 1975). Again, the spectral amplitudes should be averaged over the period band of interest. Measurements at shorter periods may be necessary to obtain measurements for close-in data on small explosions. Log A instead of Log $^{A}/_{T}$ should be used in the magnitude calculation.

Distance Normalization. With spectral amplitudes, the distance normalization term must account for geometric spreading and Q. For a first approximation, average Q for continental paths could be used. The normalization term can be refined for each station by evaluating the group velocity data obtained in the development of phase matched filters.

Corrections. The comments on station magnitude corrections for m_b apply to M_S. Reliable corrections for evaluating azimuthal variations in amplitude are a must for M_S. Phase matched filters appear to provide a means for very precise determinations of the relative magnitudes for explosions within a test site.

References

1 - Herrin E. T., 1976 "Improved Method for Detection of Long-Period Rayleigh Waves for Identification of Earthquakes and Underground Explosions, "Semi-Annual Technical Report to AFOSR, ARPA-AFOSR Contract AF 44620-73-C-004.

2 - Marshall, P. D., and P. W. Basham, "Discrimination Between Earthquakes and Underground Explosions Employing an Improved M_S Scale," Geophysical J. R. Astr., Soc 28, 432-458, 1972.

3 - Mitchell, B. J., "Regional Rayleigh Wave Attenuation in North America, J. Geophs. Res., 80, 4904-4916, 1975.

California Institute of Technology, Pasadena

Definition and Estimation of Surface Wave Magnitudes

David G. Harkrider

Standard time domain estimates of M_s using 20 second period arrivals should be avoided when they appear within a cycle of the Airy phase or when they follow it. More work should be done on the use of the Airy phase as an estimate of M_s, especially at large distances. For African events measured on the African continent and for regions where the Gutenberg continental modal is appropriate, the 20 second arrival is very small compared to the 10-18 sec arrivals which form the inverse or short period branch of the observed and theoretical Rayleigh wave group velocity curve.

Harkrider (1977) and Rodi et al. (1978) have tried to reduce the scatter in surface wave amplitude estimates of $\Psi(\infty)$ from Sahara events by inverting for propagation structure for each observation path. The recorded Love waves are used to estimate the Tectonic correction.

A similar study for NTS events using Albuquerque and Tucson recordings was done by Bache et al. (1978). They felt that the good fits to observations were suspect because of the special $\gamma(\omega)$'s, spectral attenuation factors, which were used to obtain the agreement. Some of the $\gamma(\omega)$'s require rather extreme depth variations in Q. The validity of the $\gamma(\omega)$ models must be determined before this method can be considered a viable approach.

References

Bache, T. C., W. L. Rodi, and B. F. Mason (1978), Source Amplitudes of NTS Explosions Inferred from Rayleigh Waves at Albuquerque and Tucson, Systems, Science and Software Topical Report SSS-R-78-3690.

Harkrider, David G. (1977), Preliminary Report on the French Sahara Event Saphire, Semi-Annual Technical Report 1 April 1977-30 September 1977, California Institute of Technology, AFOSR #F49620-77-C-0022.

Rodi, W. L., J. M. Savino, T. G. Barker, S. M. Day, and T. C. Bache (1978), Analyses of Explosion Generated Surface Waves in Africa, Results from Discrimination Experiment and Summary of Current Research, Systems, Science and Software Quarterly Technical Report SSS-R-78-3653.

8. DEFINITION OF SURFACE WAVE MAGNITUDE

by

Thomas C. Bache

Systems, Science and Software
P. O. Box 1620
La Jolla, California 92038

Surface wave magnitude, as conventionally defined, is based on some time domain measurement of amplitude and period. Much work has been done to empirically determine the "best" distance corrections, corrections for dispersion effects, etc. (e.g., Marshall and Basham, 1972). This is important, but perhaps too little attention has been given to development of more consistent methods for measuring the amplitude used in these formulas.

For yield estimation the amplitude we want to measure is the spectral amplitude of the source. If the dispersion were always about the same, conventional measurement techniques would give a consistent estimate for this and would work very well.

Perhaps the best way to correct surface wave amplitude for path differences is to directly model the path. This is the approach taken by Bache, Rodi and Harkrider (1978) which is described in some detail under Topics 10 and 13. This approach assumes that a path model that correctly models the dispersion also gives an accurate representation for the path amplification. This procedure takes some time and effort, but the important paths could be modeled over a period of time and surface wave amplitude measurements could then be accurately corrected.

Let us return to the point that spectral amplitude is the quantity of interest. Under Topic 1 I describe an automated

method for determining a spectral body wave magnitude. Essentially the same procedure will work quite nicely for surface waves. Let us call the result $\hat{M}_s$, following the notation $\hat{m}_b$ used for body waves.

To compute $\hat{M}_s$ we follow the procedure outlined under Topic 1 for $\hat{m}_b$, after some modification to account for the dispersed nature of the signal. The $\hat{m}_b$ is based on seismogram processing with the MARS signal analysis program. This program is used for P wave detection and for calculation of a discriminant based on short period P waves. It is also an excellent tool for determining surface wave dispersion and is often used for that purpose.

The P wave detection algorithm in MARS searches for an undispersed arrival and computes the amplitude spectrum of that arrival. To implement this algorithm for surface waves we need to have a prior estimate of the dispersion for the path. "Undispersing" the signal with this information, we then compute $\hat{M}_s$ with nearly the same algorithm used for $\hat{m}_b$.

The algorithm we have outlined is now being implemented and tested at Systems, Science and Software (S^3). We expect that the impressive results achieved for body waves will be matched for surface waves. We are, perhaps, premature in including this in a state-of-the-art assessment of capabilities. But if $\hat{M}_s$ works nearly as well as we expect, it represents a capability to sharply increase the accuracy and consistency of our measurements of surface wave amplitude.

REFERENCES

Bache, T. C., W. L. Rodi and D. G. Harkrider (1978), "Crustal Structures Inferred from Rayleigh-Wave Signatures of NTS Explosions," BSSA, 68, pp. 1399-1413.

Marshall, P. D. and P. W. Basham (1972), "Discrimination Between Earthquakes and Underground Explosions Employing an Improved M_s Scale," Geophys. J., 28, pp. 431-458.

Teledyne Geotech
Robert R. Blandford

Definition of Surface Wave Magnitude

von Seggern (1977) has defined a spectral magnitude as the noise-corrected spectral amplitude between 20 and 45 seconds. He also derived the appropriate distance amplitude relation required in order to allow for the fact that the spectral measurement removes the effects of dispersion. The technique was compared to analyst measurements on NORSAR long-period signals and was shown to have a slightly smaller variance; and good correlation was found with analyst M_s values for 100 events recorded at ANMO. The technique was developed in order to have an "automatic" method suitable for NEP.

von Seggern, D. W., (1977). Methods of automating routine analysis tasks in perparing a global seismic bulletin, TR-77-13, Teledyne Geotech, Alexandria Virginia.

Estimation of Surface Wave Magnitude

von Seggern, (1977) has used a common well-distributed network of 38 teleseismic WWSSN stations to compute M_s magnitudes of Boxcar and Milrow. Boxcar has a magnitude 0.55 M_s units higher than Milrow even though the shot point media are well matched. A result of this sort, unless explained, throws a serious cloud over any attempt to use Rayleigh waves in a threshold test ban treaty. A possible explanation is that Amchitka, even though it lies on an island arc, may have an oceanic structure so far as Rayleigh wave generation of 20 second waves in concerned. Harkrider's theory shows that a shot in an oceanic crust will have an M_s 0.35 units lower than a shot in the Basin and Range. A satisfactory resolution of this problem will require substantial research.

As far as relative yields within a single test site are concerned, however, Rayleigh waves appear to offer a substanital advantage. Rivers and von Seggern (1978) have shown that the fluctuations in M_s across arrays are approximately 1/3 those for m_b so that by reciprocity; relative yields determined by one station using M_s should be as good as relative yields determined by nine stations using m_b. Rivers and von Seggern also confirmed Mack's (1972) result that scatter is less along the raypath than perpendicular to it, so that relative yields are best determined by stations on the line between two events than by stations perpendicular to that line.

Mack, H. (1972). Spatial coherence of surface waves, Report No. SAAC-8, Teledyne Geotech, Alexandria, Virginia.

Rivers, D. W., and D. H. von Seggern, (1978). Random scattering effects on Rayleigh-wave amplitudes and phases, TR-78-3, Teledyne Geotech, Alexandria, Virginia.

von Seggern, D. W., (1977). Intersite magnitude-yield bias exemplified by the underground nuclear explosions Milrow and Boxcar, TR-77-3, Teledyne Geotech, Alexandria, Virginia.

Pennsylvania State University, University Park
Shelton S. Alexander

ESTIMATION OF SURFACE WAVE MAGNITUDE

In recent years several important advances have been made in surface wave analysis techniques that permit improved surface wave estimates of source moment (magnitude) for both explosions and earthquakes. Significant breakthroughs in long-period and broad-band seismic instrumentation, coupled with high dynamic range digital recording and advanced digital processing methods, now put us in the position to implement these analysis techniques for routine use and test them on large data sets.

By far most of the Ms estimates to date for explosions (and earthquakes) have been based on the original Gutenberg approach of measuring the amplitude (time domain) of the 20 second period Rayleigh wave signal and applying a distance connection that incorporates geometrical spreading, dispersion, and attenuation. Station corrections, empirically derived or theoretically calculated, can be used to improve Ms estimates following this approach, but there remains the inherent problem of using a time-domain measurement to obtain a spectral estimate from a dispersed wave-form. Further, only the excitation at a single period (~20 sec) is used as contrasted with a broader frequency band estimate. The routine application of (better) time domain and spectral methods has been limited, because until recently only analog recordings were available for most stations and analog to digital conversion was a time-consuming, tedious process; however, at the present time the existing and planned global digitally recording stations have effectively removed this obstacle for future Ms determinations.

Despite these difficulties Marshall, et al. (1971) have shown that there is small scatter about the relationship $Ms = \log Y + 2.0$ for explosions in consolidated rock (tuff, salt, granite, andesite, and sandstone) ranging in yield (Y) from 4 tp 1200 kilotons. They applied a path connection to account for differences in amplitude caused by lateral variations in dispersion. They report a standard deviation (after correction) of .25 for a single observation and a standard deviation of .1 in the mean Ms value. This implies that yield estimates from conventional Ms determinations will be in error by less than a factor of 2 (assuming one or more calibration events with known yield are available for a given source region); there remains a question of the universality of magnitude-yield relationships for both surface waves and body waves. Von Seggern (1973) further shows that scatter in Ms estimates is only half that for m_b.

Thus, for cases in which several stations with good signal to noise ratings can be used to estimate Ms, reliable yield estimates should be obtained. (An important exception must be made here, however, for events with associated thrust or dip-slip type tectonic effects, as discussed later in this section.)

One significant general limitation to using Ms for yield determinations is the higher limit in magnitude (compared to teleseismic P-waves) at which adequate signal to noise levels for reliable signal amplitude measurements can be attained in the conventional approach to determine Ms. Typically, then, at small yields the number of stations, azimuthal coverage, and S/N ratios become small and substantial uncertainties must be assigned to the average Ms values.

A number of techniques have been developed to improve the S/N for surface waves, even for single station observations. For example, Alexander and Rabenstine (1967 a, b) developed a matched filter method for single station or array enhancement of dispersed surface waves. In effect all of the dispersed long-period energy is compressed into a pulse-like waveform while the rms noise level remains unchanged; single station enhancements of 6-9 db are typical using this approach, and array sums add on additional $\sqrt{n}$ for n stations. Herrin and Goforth (1977) developed a phased matched filter technique to eliminate effects of multipathing. Polarization filtering (e.g. Simons, 1968; Choy and McCamy 1973; and Smart 1977) that takes advantage of the $\pi/2$ phase shift between the vertical and radical components of Rayleigh wave signals also can help to enhance surface wave signals and in principle can be combined with matched filtering and array processing to achieve further gains. Finally frequency-wave number analysis using arrays such as LASA, ALPA, or NORSAR can further improve S/N for small events and separate them from overlapping events (or noise) propagating in different directions across the array matched filtering in combination with f-k analysis can further enhance weak signals from particular areas of interest.

Spectral methods for surface wave analysis are likely to provide the most powerful and complete description of both explosion and earthquake sources. Theoretical analysis of surface waves are typically developed in the frequency (wave-number) domain where for each mode, component effects (source spectrum, transfer function, instrument response) all appear as single products in contrast to the time domain where these effects are convolved together. Correction for path and instrument effects is considerably more straightforward and less ambiguous in the frequency domain than in the time domain. Furthermore the use of spectral estimates eliminates the significant amplitude corrections for dispersion that are needed for time domain estimates of Ms. Techniques for obtaining good surface wave spectral estimates include: simple fourier transforms of signal windows; narrow band filtering; combined matched filtering and fourier transforms for single stations or arrays (Alexander and Lambert, 1971); phased matched filtering (Herrin and Goforth, 1977); and frequency-wave number analysis for arrays (e.g. Smart 1977). Through these approaches useful spectral estimates can commonly be obtained for weak signals for which conventional Ms determinations are very poor. More work on spectral magnitude estimates is needed, however.

Finally, it should be emphasized that short-period Ms estimates from Lg signals provide another means of estimating yield. Although studies of Lg magnitude vs yield relationships are less complete than for long-period surface waves, there is evidence that suggests that Lg estimates may be as good as or possibly better than Ms or M_b for yield determinations. (Baker 1970 and Blandford 1979, among others, report smaller variances for Lg magnitudes than for M_b.) It is commonly argued that propagation and station effects make the use of Lg for magnitude (yield) estimation suspect. However, recent work at Penn State indicates that for any given source region and a common network of stations, propagation effects and station terms can be determined, and in turn reliable source strength (moment) estimates can be made for each individual event.

Clearly there are limitations to the use of Lg, probably the most significant being that Lg does not survive transmission across continental margins. On the other hand at regional distances Lg commonly is the strongest signal present and consequently will provide the lowest detection threshold and the best S/N for yield estimation for small events. A lot remains to be done to understand Lg generation and propagation, but the use of empirical Lg vs yield or Lg vs M_b curves need not await these studies.

Further studies needed to improve Ms estimates of yield include the following:

1. Multipathing

McGarr (1969a), Capon (1970), and Sobel and von Seggern (1978) have all pointed out the very serious scatter in surface-wave amplitudes that can arise due to multipathing, focusing, and refraction of surface waves. In general, these effects can become more severe with epicentral distance and are especially troublesome when major tectonic boundaries lie on the travel path.

Multipathing with significant time delays of the type studied by Capon (1970) can be rather easily dealt with. Its effect is often clear on the seismogram itself or in the spectrum where "scalloping appears." Tsai (19) has shown that application of homomorphic filtering can reduce the seismogram to a direct arrival only when the multipath arrivals are sufficiently separated. More recently, Herrin and Goforth (1977) have developed a method for eliminating multipath arrivals of small time delay and recovering the direct surface wave. It is necessary now to use these techniques on suites of seismograms from different explosions to find out whether this processing will reduce the scatter in Ms estimates and improve yield determination.

2. Effects of Major Boundaries on Surface-Wave Amplitude

For waves traversing major structural boundaries, there will be a readjustment of the model response and a consequent change in amplitude over the entire spectrum. McGarr (1969b) gave some of the first clear empirical evidence of this effect. McGarr and Alsop (1967) have given approximate analytic treatments of the problem. Drake (1972) has used a finite element method to study the problem. This latter approach yields excellent results and is capable of modelling complex transition zones.

The above studies refer to normal incidence of the traveling waves to the boundary. In such cases the reflection coefficients of long-period surface waves are small. The problem of oblique incidence was recently treated by Chen and Alsop (1979) who showed that, even for oblique incidence to roughly 45°, the reflection coefficients should be negligible.

3. Surface-Wave Coupling at the Source

Usually the long-period surface waves from explosions are synthesized using a plane-layered earth model identical for source, path, and receiver structures. But the partitioning of energy at the source depends critically on the exact close-in structure which may differ significantly from the simplified models used in the synthesis. Bache et al. (1978) have attempted to split the source from the path by generating the amplitude response in a source structure and the dispersion relation in a different path structure.

There is a need for a more precise approach to the synthesis problem using extensive finite element models to propagate the waves away from the elastic radius of the explosion. Viccelli (1973) has already experimented with 2-D models for explosive wave generation, but the grid sizes need to be considerably enlarged to accomodate waves from 1 to 20 sec periods and distances out to several tens of km. in order to accurately assess Rayleigh excitation in laterally inhomogeneous media. Furthermore, 3-D modelling will be required to study source structures which are not axisymmetric such as the Amchitka Island test site. Schlue (1979) has presented the formalism for 3-D finite-element modelling, and further consideration should be given to utilizing this type of approach.

4. Resolution of the Combined Explosion-Strain Release Mechanism

The work of Toksoz and Kehrer (1972) in defining the strain release of explosions at various U.S. and USSR test sites is limited by the assumption of pure strike-slip motion on a vertical fault. This is true also of Harkrider's (1978) recent study of a Sahara explosion. A more general source inversion is required, and the formalism of moment-tensor inversion introduced by Mendiguren (1977) provides a powerful tool. This technique should be utilized on recent shots where high-quality digital data is available from SRO, HGLP, and array stations. Results can be used to indicate the probable contamination of both spectral-domain amplitude and time-domain amplitude estimates of Ms for explosions of interest. An enhancement of the technique has already been provided by Patton (1978) who uses a reference event to remove path effects. If a reference explosion relatively free of tectonic strain release is available, then the composite mechanism of each nearby explosion can probably be determined with good accuracy, permitting a "true" Ms to be determined for all the explosions.

(See also the following discussion of tectonic release effects.)

2. <u>Use theoretical Love to Rayleigh excitation estimates for dominantly thrust (or dip-slip) focal mechanisms to assess the viability of correcting Ms estimates for tectonic effects</u>. The approach would be to use observed Love wave amplitudes vs azimuth together with theoretical L/R ratios vs azimuth to estimate the tectonic contribution to observed Rayleigh wave amplitudes. Another possible approach to accomplish the same objective would be to carry out a moment-tensor inversion (preferably in the frequency domain) to estimate the explosion vs tectonic contribution. The key to success of either of these approaches will be to include Love-wave observations.

3. <u>Use data from subsets of explosions with the smallest tectonic effects to establish an asymtotic, empirical Ms vs M_b relationship</u>. If perturbations in Ms from this curve correlate well with observed Love wave excitation, then an empirical correction procedure could be developed. However, this approach will fail if the tectonic contribution to Rayleigh waves overcomes the explosion component and changes the polarity of the observed Rayleigh signals, as sometimes is the case in thrust-type tectonic settings. Also this approach assumes that M_b estimates will be unaffected by tectonic contributions, and this question needs to be studied further. For example, tectonic contributions, from thrust-type tectonic "release" will be in the sense to increase teleseismic P-wave amplitudes in all or nearly all azimuths, if the origin time is not significantly delayed with respect to the explosion time; if significant delays are present, they should be manifested in P-wave spectral interference effects (and also azimuthal variations in the interference pattern if the "epicenter" for the tectonic release is displaced from the explosion).

4. <u>Investigate the stability of Lg magnitude estimates vs yield or vs M_b</u>. There is accumulating a body of evidence that Lg excitation is relatively insensitive to tectonic "release" as manifested by long-period fundamental mode Love and Rayleigh waves from NTS and North African explosions. Furthermore, the observed scatter in Lg magnitude vs yield at single stations is comparable to or smaller than for M_b. This stability needs to be investigated for other tectonic settings, particularly ones characterized by thrust or dip-slip mechanisms. If this stability holds for all or most types of tectonic source regions, Lg observations can provide reliable additional explosion yield estimates for any source area. The magnitude threshold for useful Lg measurements, especially at regional distances, is comparable to or better than that for body waves, provided reasonably uncomplicated continental propagation paths can be used. It is commonly argued that propagation and station effects make the use of Lg for magnitude (yield) estimation suspect. However, recent work that we have done indicates that for any given source region and a common network of stations, propagation effects (attenuation vs distance) and station terms can be estimated, and in turn that reliable source strength (moment) determinations can be made.

REFERENCES

Alexander, S.S. and Rabenstine, D.B. (1967a) Detection of surface waves from small events at teleseismic distances, Seismic Data Laboratory Report No. 175, Teledyne Geotech.

Alexander, S.S. and Rabenstine, D.B. (1967b) Rayleigh wave signal to noise enhancement for a small teleseism using LASA, LRSM, and observatory stations, Seismic Data Laboratory Report No. 194, Teledyne Geotech.

Alexander, S.S. and J.W. Lambert (1971) Single-station and array methods for improved surface wave spectral estimates, SDL Report No. 264, 104 p.

Alewine, (1972) Theoretical and Observed Distance Corrections for Rayleigh-waves Magnitude, BSSA 62, 1611 - 1619.

Anglin (1971) Discrimination of Earthquake and Explosions Using Short Period Seismic Array Data, Nature 233, 51 - 52.

Anglin and Israelson (1973) Discrimination of Earthquakes and Explosions Using Multi-Station Seismic Data, BSSA 63, 321 - 323.

Bache, T.C., W.L. Rodi and D.G. Harkrider (1978) Crustal structures inferred from Rayleigh-wave signatures of NTS explosions, BSSA, 68, 1399 - 1414.

Basham (1969) Canadian Magnitudes of Earthquakes and Nuclear Explosions in South Western North America, Geophys. J. 17, 1 - 13.

Basham and Anglin (1973) Multiple Discriminant Screening Procedure for Test Ban Verification, Nature 246, 474 - 476.

Bolt (1976) Nuclear Explosions and Earthquakes - The Partial Veil. W.H. Freeman, San Francisco, CA, p. 151 - 168.

Booth, Marshall and Young (1975) Long and Short Period Amplitudes of Earthquakes in the Range 0° - 114°, Geophys. J. 39, 523 - 538.

Bruno, Espinoza and Oliver (1963) Relative Excitation of Surface Waves by Earthquakes and Underground Explosions in the California-Nevada Region, J.G.R. 68, 3501 - 3513.

Bungum and Tjostheim (1976) Discrimination Between Eurasian Earthquakes and Underground Explosions Using the M_b:Ms Method and Short Period Autoregressive Parameters, A. Geophys. J. 45, 371 - 372.

Capon, J. (1970) Analysis of Rayleigh-wave multipath propagation at LASA, BSSA, 62, 1259 - 1268.

Carder, Gordon and Jordan (1963) Seismic Waves from an Underground Nuclear Explosion Under a Desert Valley (Mississippi Event), USGGS Report, June, 1963.

Chen, T.C. and L.E. Alsop (1979) Reflection and transmission of obliquely incident Rayleigh waves at a vertical discontinuity between two welded quarter-spaces, BSSA, 69, 1409 - 1424.

Choy, G. and K. McCamy (1973) Enhancement of long period signals by time varying adaptive filters, J. Geophys. Res., 78, 3503 - 3511.

Dahlman, Israelson, Austegard and Hornstrom (1974) Definition and Identification of Seismic Events in the USSR in 1971, BSSA, 64, 607 - 636.

Dahlman and Israelson (1977) Monitoring Underground Nuclear Explosions, Elsevier, p. 219 - 235.

Der (1973) Ms - M_b Characteristics of Earthquakes in the Eastern Himalayan Region, Geotech Report No. 296AD, 759 - 835, Teledyne Geotech, Alexandria, Virginia.

Douglas, Hudson and Kembhavi (1971) The Relative Excitation of Seismic Surface and Body Waves by a Point Source, Geophys. J. 23, 451.

Drake, L.A. (1972) Rayleigh waves at a boundary by the finite element method, BSSA, 62, 1259 - 1268.

Elvers (1975) Seismic Event Identification by the M_b (Ms) Method. National Defense Research Institute, Stockhol[illegible] [illegible]ort C20070 - T1.

Ericsson (1971) Identificatio[illegible] [illegible]nderground Nuclear Explosions and Earthquakes. FOA reports V. [illegible], No. 8, Research Institute of the Swedish National Defense.

Errasson (1970) Event Identification for Test Ban Control, BSSA 60, 1521 - 1546.

Evernden (1969) Identification of Earthquakes and Explosives by use of Teleseismic Data, J.G.R. 74, 828 - 3856.

Evernden and Filson (1971) Regional Dependence of Surface Wave vs. Body Wave Magnitudes, J.G.R. 76, 3303 - 3308.

Evernden (1975) Further Studies on Seismic Discrimination, BSSA 65, 359 - 392.

Evernden, Best, Pomeroy, McEvilly, Savino and Sykes (1971) Discrimination Between Small Magnitude Earthquakes and Explosions, J.G.R. 76, 8042 - 8055.

Filson and Bungum (1972) Initial Discrimination Results from the Norwegian Seismic Array, Geophys. J. 31, 315 - 328.

Forsythe (1976) Higher Mode Rayleigh Waves as an Aid to Seismic Discrimination, BSSA 66, 827 - 841.

Harkrider (1978) Report on Sahara Shot.

Hasegawa (1973) Sur ace and Body-wave Spectra of CANNIKIN and Shallow Aleutian Earthquakes, BSSA, 1201 - 1226.

Herrin, E. and T. Goforth (1977) Phase-matched filters: Applications to the study of Rayleigh waves, BSSA, 67, 1259 - 1277.

Israelson and Yamamoto (1974) Seismic Magnitudes from Explosions and Earthquakes Obtained at the Hagfors and Matsushiro Observatories. National Defense Research Institute, Stockholm, Report C20010 - T1.

Kaila (1970) Decay Rate of P-wave Amplitudes from Muclear Explosions and the Magnitude Relations in the Epicentral Distance Range $1^{o} - 98^{o}$, BSSA 60, 447 - 460

Kalar and Pruvost (1975) Earthquake Simulation by Nuclear Explosions, Nature 253, 242 - 245.

Landers (1972) Some Interesting Central Asian Events in the Ms: M_b Diagram. Geophys. J. 31, 329 - 339.

Lieberman and Pomeroy (1969) Relative Excitation of Surface Waves by Earthquakes and Underground Explosions, J.G.R. 74, 1575 - 1590.

Marshall Ms vs yield-call Blandford.

Marshall, P.D., A. Douglas and J.A. Hudson (1971) Surface Waves from Underground Explosions, Nature 234, No. 5323, 8 - 9.

Marshall and Basham (1972) Discrimination Between Earthquakes and Underground Explosions Employing an Improved MsScale. Geophys. J. 28, 431 - 458.

Marshall and Key (1973) An Analysis of Seismic Waves from Earthquakes and Explosions in the Sino-Soviet Area During 1966. UK Atomic Energy Authority, AWRE, Aldermaston Report No.) 5173.

Marshall, Springer and Rodean (1979) Magnitude Corrections for Attenuation in the Upper Mantle. Geophys. J. 57, 609 - 638.

McEvilly and Peppin (1972) Source Characteristics of Earthquakes, Explosion and After Events, Geophys. J. 31, 67 - 82.

McGarr, A. (1969a) Amplitude Variations of Rayleigh waves - Horizontal Refractions, BSSA, 59, 1281 - 1306

McGarr, A. (1969b) Amplitude Variations of Rayleigh waves - Propagation across a Continental Margin, BSSA, 59, 1281 - 1306.

McGarr, A. and L.E. Alsop (1967) Transmission and Reflection of Rayleigh Waves at Vertical Boundaries, J.G.R. 72, 2169 - 2180.

Mendiguren, J.A. (1977) Inversion of Surface Wave Data in Source Mechanism Studies, J.G.R. 82, 889 - 894.

Mitchell, B.J. Anelasticity of the Crustal Upper Mantle Beneath the Pacific Ocean from Inversion of Observed Surface Wave Attenuation, Geophys. J. 46, 521 - 533.

Mitchell et al. (1976) Attenuation of Love and Rayleigh Waves Across the Pacific at Periods Between 15 and 110 seconds, BSSA 66, 1189 - 1202.

Mitchell, B.J., N.K. Jacoub and A.M. Correig (1977) A Summary of Seismic Surface Wave Attenuation and its Regional Variation across Continents and Oceans, in Geophysical Monograph 20, The Earth's Crust, J.G. Heacock, ed., pp. 405 - 425, AGU, Washington, D.C.

Nuttli and Kim (1975) Surface Wave Magnitudes of Eurasian Earthquakes and Explosions, BSSA 65, 693.

Pasechnik, Daskkov, Polikerpova and Gamburtseva (1970) The Magnitude Method for Identification of Underground Nuclear Explosions. 1 gv. Akad Nauk USSR, Earth Physics 1, 28 - 36.

Patton, H. (1978) Source and Propagation Effects of Rayleigh Waves from Central Asia Earthquakes, Ph.D. Thesis, MIT, Cambridge, MA.

Peppin and McEvilly (1974) Discrimination Among Small Magnitude Events on Nevada Test Site, Geophys. J. 37, 227 - 243.

Press, Delvart and Gilman (1963) A Study of Diagnostic Techniques for Identifying Earthquakes. J.G.R. 68, 2909 - 2928.

Ringdal (1976) Maximum Likelihood Estimation of Seismic Magnitude, BSSA 66, 789 - 802.

Savino, Sykes, Lieberman and Molnar (1971) Excitation of Seismic Surface Waves with Periods of 15 - 70 sec. for Earthquakes and Underground Explosions. J.G.R. 76, 8003 - 8020

Schlue, J.W. (1979) Finite Element Matrices for Seismic Surface Waves in Three-dimensional Structures, BSSA 79, 1425 - 1438.

Shapira and Kulhanek (1978) Conventional and Spectral Short-Period Body Wave Magnitudes. BSSA 68, 1195 - 1198.

Simon, R. (1968) PHILTRE-A Surface Wave Particle Motion Discrimination Process, BSSA 58, 629 - 637.

Sipri (1968) Seismic Methods of Monitoring Underground Explosions, International Institute for Peace and Conflict Research, Stockholm.

Smart, E. (1977) A Three-Component Single-Station Maximum Likelihood Surface Wave Processor, SDAC-TR-77-14, Teledyne Geotech, Alex., Va.

Sobel, P.A. and D.H. von Seggern (1978) Applications for Surface Wave Ray Tracing, BSSA 68, 1359 - 1380.

Tatham, Forsyth and Sykes (1976) The Occurrence of Anomalous Seismic Events in Eastern Tibet. Geophys. J. 45, 451 - 481.

Thomas, Marshall and Douglas (1978) Rayleigh Wave Amplitudes from Earthquakes in the Range 0^{o} - 150^{o}, Geophys. J. 53, 191 - 200.

Toksoz, M.N. and H.H. Kehrer (1972) Tectonic Strain Release by Underground Nuclear Explosions and its Effect on Seismic Discrimination, GHRAS 31, 141 - 161.

Tsai and Aki (1971) Amplitude Spectra of Surface Waves from Small Earthquakes and Underground Nuclear Explosions.

Viccelli, J.A. (1978) Spallation and the Generation of Surface Wave by an Underground Explosion, J.G.R. 78, 2475 - 2487.

von Seggern, D.H. (1973) Joint Magnitude Determination and Analysis of Variance for Explosion Magnitude Estimates, BSSA 63, 827 - 853.

Ward and Toksoz (1971) Cause of Regional Variation of Magnitudes, BSSA 61, 649 - 670.

Willis, DeNoyer and Wilson (1963) Differentiation of Earthquakes and Underground Nuclear Explosions on the Basis of Amplitude Characteristics, BSSA 53, 979 - 987.

Yacoub, R.K. and B.J. Mitchell (1977) Attenuation of Rayleigh Wave Amplitudes across Eurasia, BSSA 67, 751 - 769.

13. EFFECTS OF SOURCE REGION PROPERTIES ON SURFACE WAVE GENERATION

by

Thomas C. Bache

Systems, Science and Software
P. O. Box 1620
La Jolla, California 92038

Introduction

There has been a great deal of research on the generation and propagation of surface waves. Techniques for computing synthetic surface wave seismograms are widely used. Generally this involves an elastodynamic source representation (e.g., multipolar expansion, moment tensor) together with plane-layered earth models. Considerable progress has been made in finding plane-layered earth models that give dispersion in close agreement with that observed. The major problems that remain are:

- What is the proper source representation?
- What is the effect of nonplanar layering and lateral heterogeneities?
- What is the proper Q model?

Our response to the ARPA request for a state-of-the-art 34
assessment of the use of surface waves for seismic yield determination is mainly concerned with assessing the current knowledge of the answer to these questions. Under Topic 10, we discuss the dependence of the surface wave coupling on the near-source nonlinear region properties. The tectonic contribution is discussed under Topic 11 and the effects of attenuation under Topic 12. We also present a promising new method for determining M_s under Topic 8.

In this section we discuss the general problem of Rayleigh wave generation by underground explosions. While we are attempting to assess our knowledge of the effect of the near source, linear region properties on the surface waves, we also consider corrections for the path effects.

Our focus is almost entirely on summarizing the results of two Systems, Science and Software (S^3) studies; Bache, Rodi and Harkrider (1978) and Bache, Rodi and Mason (1978). In these studies state-of-the-art techniques for determining earth structure and computing synthetic seismograms are used.

Path Models Compatible with Surface Wave Observations

The main purpose here is to discuss the influence of source region properties on surface wave generation. But the best way to test any hypothesis about this influence is to compare with actual data and this requires some means for removing path effects.

Techniques for constructing plane layered earth models from observed surface wave dispersion data are now becoming an almost routine seismological tool. An application we think especially pertinent was described by Bache, Rodi and Harkrider (1978). The main elements of that paper are:

- Rayleigh wave recordings of NTS explosions were collected from the WWSSN stations ALQ and TUC. The data were divided according to test site. It was observed that recordings from a single test site (e.g., Pahute Mesa) had very similar waveforms.

- Representative recordings were digitalized from three sites; PILEDRIVER, Yucca Flat and Pahute Mesa.

- The digitized data were processed by the S^3 MARS program and phase and group velocity curves were derived for the two paths, NTS-TUC and NTS-ALQ.

- Using generalized linear inversion techniques, plane layered earth models were found that fit the data. These models are compatible with other geophysical information about these paths.

- A Q model was constructed from western United States attenuation data collected by Mitchell (1975).

- Synthetic seismograms were computed with a simple reduced displacement potential (RDP) source. These seismograms are shown compared to typical observations in Figure 1.

This procedure ensures that the dispersion of the synthetic and observed seismograms be the same. However, the excellent waveform agreement indicates that the amplitude spectrum are fit as well.

Surface Wave Dependence on Source Material Properties

In computing the seismograms of Figure 1 there is one path model for NTS-TUC and one for NTS-ALQ. But we must deal with the fact that the local material properties are different for the three test areas. This was handled by using a two path model with the local differences accounted for in the top two kilometers of the source region model. Transition between the two is accounted for by an approximate transmission coefficient, $T(\omega)$, based on results of McGarr (1969). The equation actually used is

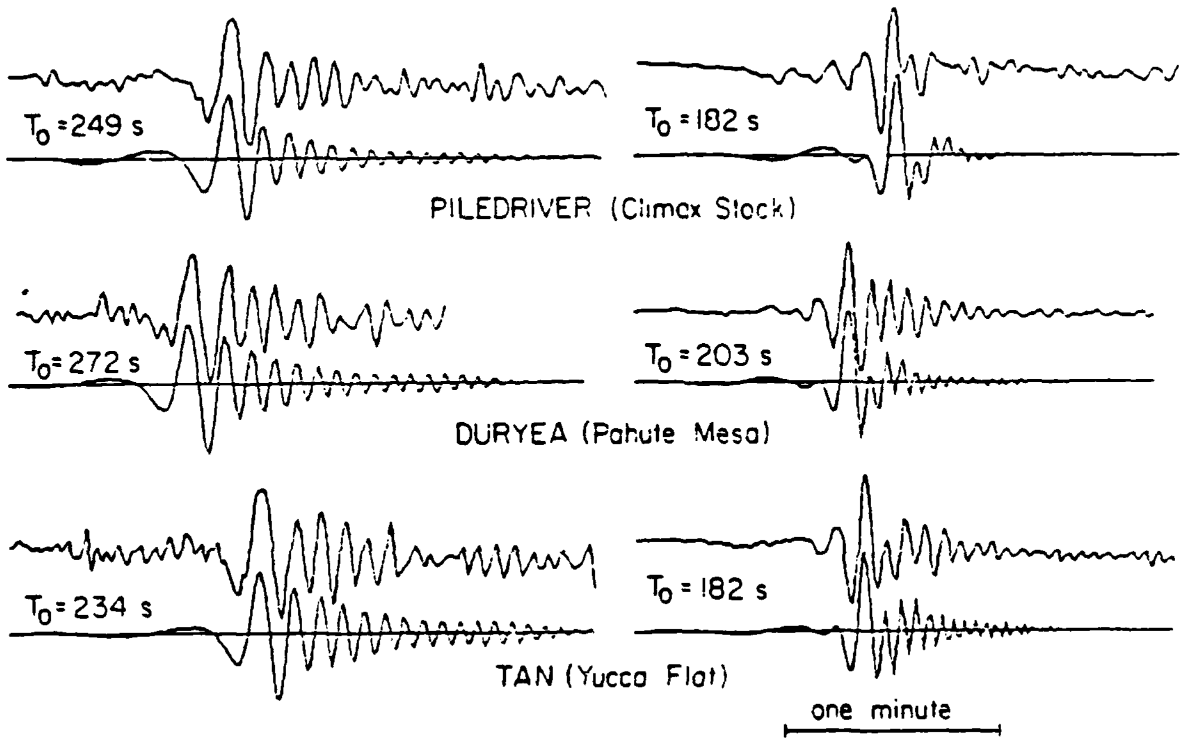


Figure 1. Theoretical and observed seismograms are compared at ALQ (left) and TUC for events in three test areas at NTS. A bar indicating one minute is shown. In each pair the observed (top) and theoretical records start at the same time with respect to the explosion detonation and this time is indicated as T_o.

$$\hat{\omega}(r,\omega) = -4\pi\mu_s \, \hat{\Psi}(\omega) \, \frac{K_{s_1}}{c_1} \, A_{R_1} \, T(\omega) \, H_0^{(2)} \left(\frac{\omega r}{c_2}\right) \, e^{-\gamma r} \left(\frac{r}{a_e \sin \Delta}\right)^{1/2},$$

where subscripts 1 and 2 indicate the source and path models.

An important question is, keeping all other factors fixed how does Rayleigh wave amplitude scale with the source material properties? The $\hat{\Psi}(\omega)$ is the RDP source and is Ψ_∞ at long period, μ_s is the shear modulus at the source, K_{s_1} is the depth dependent eigenfunction for an explosion, c is phase velocity and A_R is depth-independent amplification.

The answer to the question is shown in a direct way in Figure 2. For source regions that are not too different (e.g., Pahute Mesa and Yucca Flat) we have

$$M_s \approx \log \, (\mu_s \, \Psi_\infty) \; .$$

However, if we compare the Climax Stock to the others, we see that the relationship has a strong frequency dependence and takes no simple form.

How Adequate is an RDP Source Representation?

We have pointed out that plane-layered earth models and an RDP source are adequate to give synthetic seismograms with waveforms that match the observations. Using this comparison, we inferred the Ψ_∞ required to match the observed amplitude. This work is described in a 1978 S^3 report by Bache, Rodi and Mason.

The events considered were separated into the three populations and all were below the water table with yields between 40 and 200 KT. The inferred Ψ_∞ values are shown in Figure 3. We conclude that:

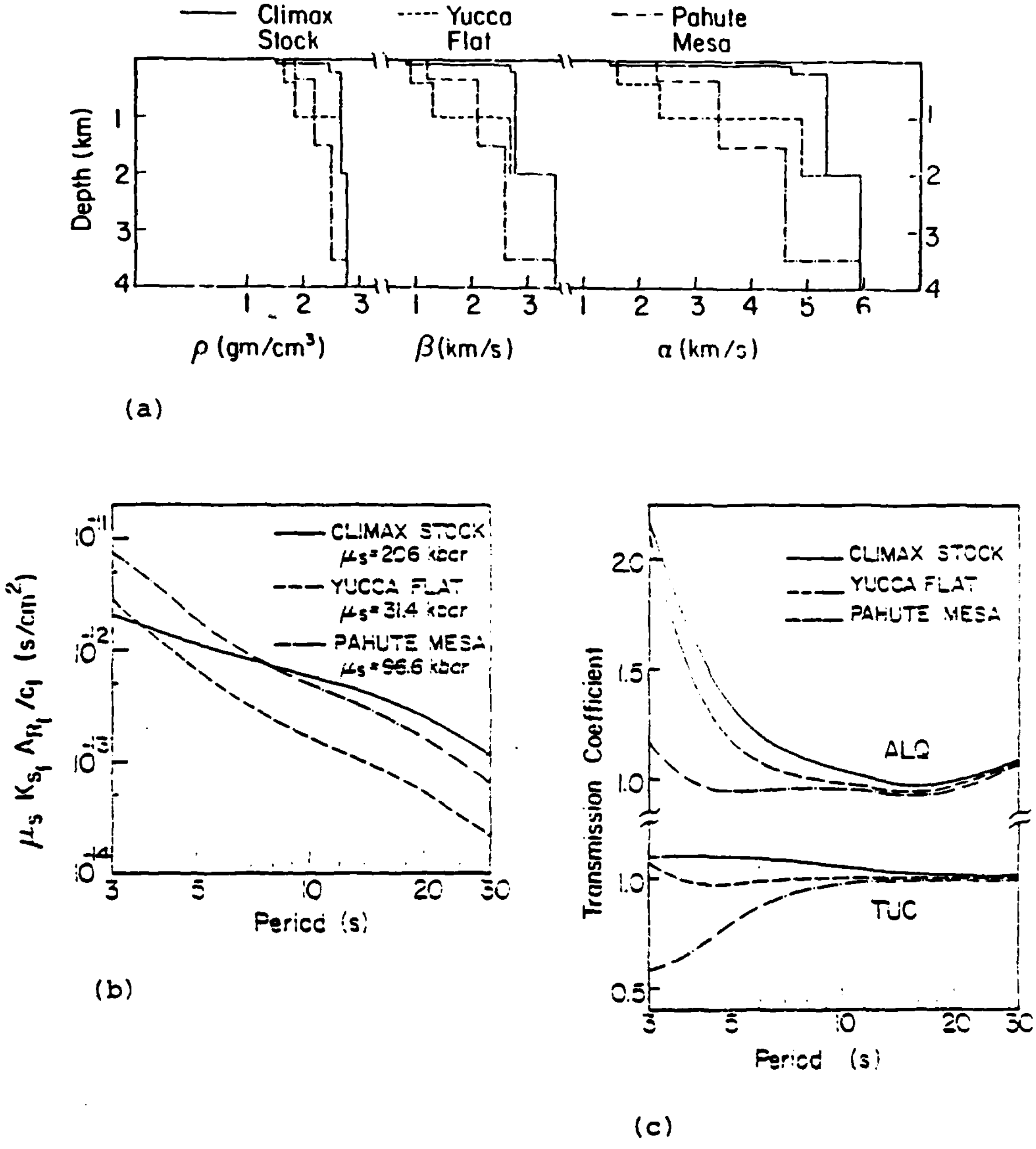


Figure 2. (a) The density, shear and compressional wave velocities are plotted versus depth for the three test areas at NTS. (b) The source amplification factor is shown for the three source areas studied. (c) The transmission coefficient $T(\omega)$ is plotted for the six source-path combinations studied.

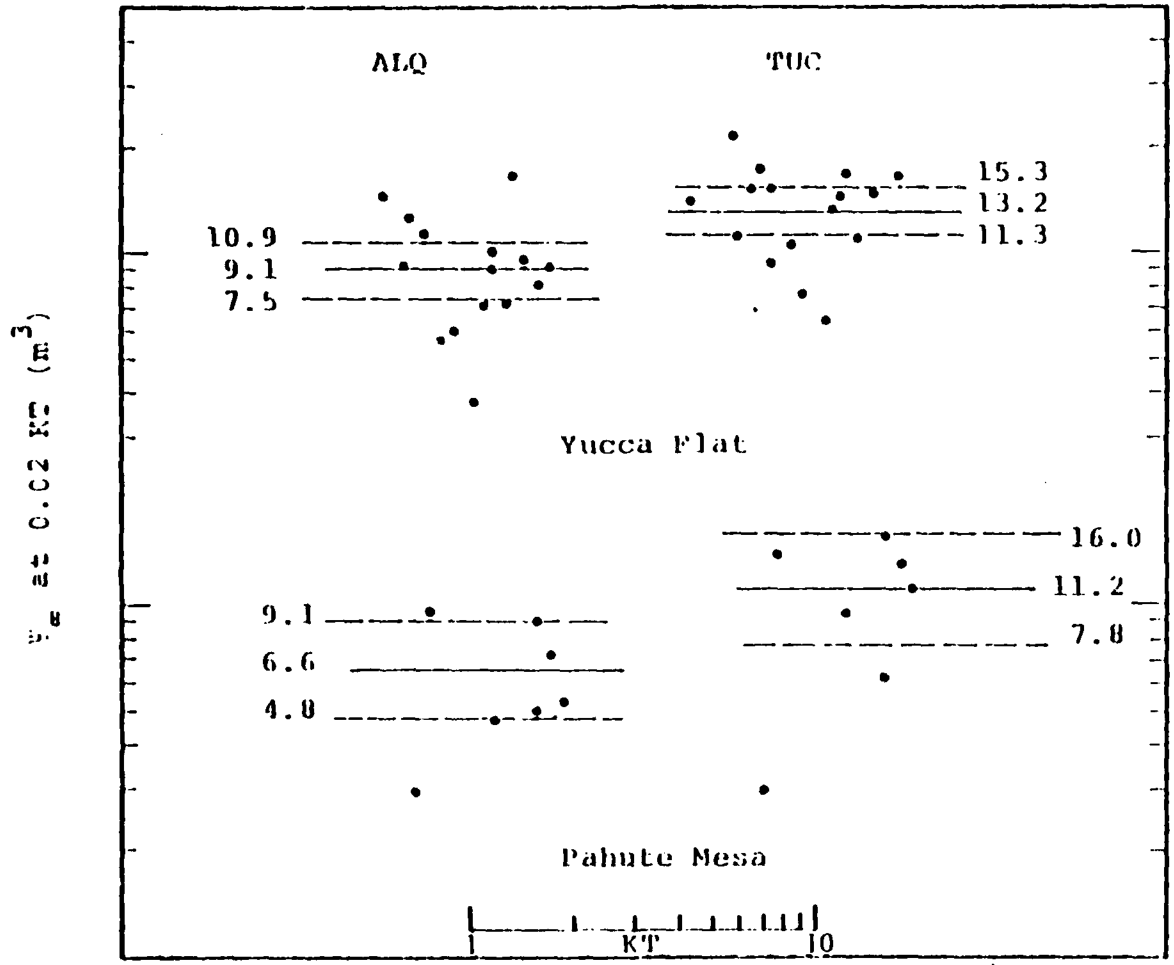


Figure 3. The Ψ_∞ values (at 0.02 KT) are plotted versus explosion yield for the Yucca Flat and Pahute Mesa events. The plot is log-log and the mean and 95 percent confidence limits on the mean are shown. One decade of yield is shown for reference.

- For each station the Ψ_∞ (scaled to a common yield) was consistent within each population — the standard deviation was about 40 percent of the mean.

- The Ψ_∞ from one station (TUC) was consistently 50 percent larger than from the other.

- The Ψ_∞ values were consistent with those derived using other methods (close-in observations, teleseismic body and surface waves, finite difference source calculations). This is discussed under Topic 10.

These events do not span a wide yield or depth range. The only depth correction made was for the depth dependence of the eigenfunctions. This turns out to be:

Yucca Flat:	$\log A \approx 0.16 \log H,$
Pahute Mesa:	$\log A \approx 0.05 \log H.$

If $H \approx W^{1/3}$, and the source function scales with yield, this translates to an M_s - log yield slope of 1.05 for Yucca Flat and 1.02 for Pahute Mesa.

In summary, the simple RDP source representation leads to reasonable results for these two stations and data spanning a narrow range in yield and burial depth. However, there are many observed phenomena that cannot be explained with an RDP source. Also, we have the ALQ-TUC amplitude discrepancy and more scatter in Ψ_∞ within each test area than we would like.

Higher Order Source Effects

Also described in the Bache, Rodi and Mason (1978) report is an attempt to delineate the influence of higher order effects. These include: double-couple generation by tectonic release, spallation, attenuation of upgoing waves by spallation and/or scattering in the near surface materials. This leads to many questions about the physics of the source.

We find that empirical estimates for spall closure (e.g., Viecelli, 1973; Sobel, 1978) suggest that the generated Rayleigh waves might be nearly the same size as those from the explosion itself. However, the spall induced Rayleigh wave is 90 degrees out of phase with that from the explosion and makes an unimportant contribution to the ALQ and TUC records for reasonable assumptions about its amplitude. We are not sure whether this result can be generalized to allow the spall effect to be ignored.

A potentially important spall related phenomenon is the attenuation of upgoing waves from the source during the spall process. This is explored by Bache, Rodi and Mason, but conclusions cannot be confidently drawn without better understanding the physics of the process. See our discussion under Topic 10 for more on this subject.

To explore the effects of a double-couple component, Bache, Rodi and Mason (1978) used the inferred double-couple solutions of Toksöz and Kehrer (1972). We find that the double-couple has almost no effect on the waveform, but simply scales the amplitude. If the events studied by Töksoz and Kehrer are typical, some 15 or 20 percent of the discrepancy between the ALQ and TUC solutions is due to the double-couple. For PILEDRIVER, their inferred double-couple dominates the solution. However, their orientation does not explain the ALQ-TUC amplitude discrepancy, but actually increases it.

Accounting for the double-couple source provides no great improvement in explaining amplitude differences between the two stations. However, its presence does significantly reduce our estimate for the explosion source amplitude (Ψ_∞), especially for PILEDRIVER. This is important because it is Ψ_∞ that is the best indicator of explosion yield.

Summary

To summarize my assessment of the current capability to model surface waves from underground explosions, I cannot improve upon the conclusions listed by Bache, Rodi and Mason (1978). These are:

- Surface wave observations of explosions can be used to infer models for the earth structure along the travel path. These models then predict synthetic seismograms that closely resemble the observed records.

- The amplitude of the source can be inferred by comparing synthetic and observed records. For Yucca Flat and Pahute Mesa events, the inferred source amplitude is in general agreement with values obtained by other methods.

- The effect of spallation was carefully explored. The spall closure impulse does not have much influence at ALQ and TUC, though it is possible that it plays a more important role in other situations. More likely to be important is the loss of energy from the upgoing waves that is associated with spallation.

- The extent to which the upgoing waves are attenuated by spallation or scattered is poorly understood. For the wavelengths of interest, it may not even occur to a significant extent. More theoretical work is required to resolve the questions. Particularly useful would be the study of Rayleigh waves from two-dimensional finite difference calculations in which spall is allowed.

- Tectonic strain release will affect the source estimates, but we may be able to correct for it if sufficient azimuthal coverage is available.

- The important PILEDRIVER event is, unfortunately, the most puzzling of those studied. The tectonic release and spall related phenomena dominate the source determination. The agreement of synthetic and observed waveforms is significantly worse for this event, suggesting that something might be unaccounted for in the vicinity of the source.

- The source level for the Yucca Flat events is best determined. A substantial proportion of the deviation from the mean Ψ_∞ for individual events is probably associated with real differences in the source coupling. It is important to know how large these differences can be for superficially identical events (individual Ψ_∞ values are given by Bache, 1978).

- The source level for Pahute Mesa events is determined with less confidence than for Yucca Flat because of larger uncertainties associated with tectonic release and spallation phenomena. Also, relatively few Pahute Mesa events were examined.

- Finally, we point out that the data for this study were taken from WWSSN film clips. This is far from ideal for many reasons, especially in the narrow yield range to which we were restricted. Digital data from well-maintained stations would be much better.

Conclusions

The capability exists to model surface waves from underground explosions in detail. The source may be of arbitrary complexity in plane layered earth models. Under Topic 10, we point out that the analytical techniques used to propagate surface waves can be linked with complex finite difference simulations of the source.

An important question is how to account for changes in the local source material when the average path properties clearly remain the same. We have been using an approximate technique that seems to give reasonable results.

When we consider the source amplitudes inferred from ALQ and TUC data using these detailed modeling procedures, we conclude that:

- Even with the simplest source model (RDP) the inferred source amplitudes are reasonably consistent.

- For similar materials, Rayleigh wave amplitude is proportional to $\mu\Psi_\infty$. For different materials (e.g., tuff and granite), the dependence on μ is much weaker and varies with period.

- Empirical estimates for spall closure suggest that the generated Rayleigh waves might be

nearly the same size as those from the explosion itself. However, spall closure appears to be unimportant for ALQ and TUC records.

Important questions that require further investigation include the following:

Path Effects

- What are the effects of lateral heterogeneities and departure from plane horizontal layering? Are such effects responsible for the 50 percent amplitude difference from ALQ to TUC?

- Attenuation (Q) is not very well-known, though this is less important, for surface waves than body waves. See Topic 12 for some further comments on this.

Source Effects

- How appropriate is the one-dimensional source representation? Known two- and three-dimensional effects include:

    1. Depth dependence of overburden pressure.
    2. Spallation and cracking to the surface.
    3. Block motion, induced fault motion, tectonic strain release.

- How are tangential (SH and Love) waves generated? Why are they generated at all azimuths? What parameters control their amplitude? How well does a single double-couple represent this perturbation on the source? Further comments on this are given under Topic 11.

- How large is spall closure? How are the free surface reflected waves in the nonlinear environment different from those in the elastic approximation?

<u>Related Questions</u>

- The ALQ and TUC study raises the question, if stations at many azimuths were used, how much would the Ψ_∞ estimate vary? How much of the "source" radiation pattern is a path effect? How much of this path effect can be corrected by careful modeling?

- What is the surface wave amplitude dependence on yield when depth is properly taken into account as in the ALQ-TUC study?

Finally, I must say that a great deal of progress has been made in understanding surface wave excitation by explosions. There are partial answers to all the questions listed above and the techniques are available to obtain more complete answers.

REFERENCES

Bache, T. C. (1978), "A Detailed Listing of the Data and Results for the Unclassified Report: Source Amplitudes of NTS Explosions Inferred from Rayleigh Waves at Albuquerque and Tucson, Systems, Science and Software Topical Report SSS-CR-78-3745 submitted to AFTAC/VSC, August, 17 pages.

Bache, T. C., W. L. Rodi and D. G. Harkrider (1978), "Crustal Structures Inferred from Rayleigh-Wave Signatures of NTS Explosions," BSSA, 68, pp. 1399-1413.

Bache, T. C., W. L. Rodi and B. F. Mason (1978), "Source Amplitudes of NTS Explosions Inferred from Rayleigh, Waves at Albuquerque and Tucson," Systems, Science and Software Report SSS-R-78-3690 submitted to ARPA/VSC, June, 91 pages.

McGarr, A. (1969), "Amplitude Variations of Rayleigh Waves — Propagation Across a Continental Margin," BSSA, 59, pp. 1281-1305.

Mitchell, B. J. (1975), "Regional Rayleigh Wave Attenuation in North America," JGR, 80, pp. 4904-4916.

Sobel, P. A. (1978), "The Effects of Spall on m_b and M_s," Teledyne Geotech Report SDAC-TR-77-12 submitted to AFTAC/VSC, April, 54 pages.

Toksöz, M. N. and H. H. Kehrer (1972), "Tectonic Strain Release by Underground Nuclear Explosions and Its Effects on Seismic Discrimination," Geophys. J., 31, pp. 141-161.

Viecelli, J. A. (1973), "Spallation and the Generation of Surface Waves by an Underground Explosion," J. Geophys. Res., 78, pp. 2475-2487.

10. SOURCE THEORY AND OBSERVATION FOR SURFACE WAVES

by

Thomas C. Bache

Systems, Science and Software
P. O. Box 1620
La Jolla, California 92038

Introduction

Under this topic I will discuss four separate subjects. These are:

1. Some experimental results on the dependence of Rayleigh wave amplitude on the properties of the source material.
2. The inference of source amplitude from surface wave recordings and comparison with other methods for determining source amplitude.
3. The development of techniques which allow consistent computation of far-field surface waves from the output of detailed finite difference calculations of the source coupling into elastic waves.
4. Some preliminary results from a study in which four detailed calculations were done of explosions in granite.

35

Experimental Determination of the Dependence of Rayleigh Wave Amplitude on Properties of the Source Material

There have been numerous studies in which Rayleigh wave amplitude, usually parameterized by M_s, is plotted versus explosion yield. Events are often separated according to characteristics of the source materials and some dependence is

sought. We have done a similar study using the Airy phase amplitudes measured on recordings at the WWSSN stations ALQ and TUC. These stations are at ranges from 700 to 900 kilometers and record small events not usually present in M_s data sets. As a disadvantage, they are off-scale for events much larger than 300 kt.

The results of this study were described in a 1977 Systems, Science and Software (S^3) Quarterly Report by Bache, Goupillaud and Mason. The Airy phase data were compared to M_s data compiled by Eisenhauer (1976). The Airy phase amplitudes were converted to M_s values using the formulas:

$$\text{ALQ:} \quad M_s^A = \log A + 2.72,$$

$$\text{TUC:} \quad M_s^A = \log A + 2.17,$$

where the constants were chosen to make M_s^A and M_s^T about the same, on the average, as Eisenhauer's M_s. An M_s was then taken to be the mean of M_s^A and M_s^T.

The data were separated by test area and a linear least square fit was made for each population when plotted versus log yield. The linear best-fit equations, together with the standard deviations, are as follows (the number of events is listed in parentheses and B is an arbitrary constant):

Pahute Mesa below the water table (9):

$$M_s = 0.86 \log W + B, \qquad \sigma = 0.09,$$

Yucca Flat below the water table (30):

$$M_s = 1.17 \log W + B - 0.93, \ \sigma = 0.19,$$

Pahute Mesa above the water table (3):

$$M_s = 1.21 \log W + B - 1.09, \ \sigma = 0.05,$$

Yucca Flat above the water table (15):

$$M_s = 0.78 \log W + B - 0.82, \sigma = 0.27,$$

Rainier Mesa, tunnel shots only (10):

$$M_s = 0.77 \log W + B - 0.41, \sigma = 0.19.$$

The M_s data together with the best fitting lines are plotted in Figure 1. As expected from the large σ values for some populations, there is considerable scatter in the data.

There are seven Pahute Mesa and eleven Yucca Flat events for which the Airy phase M_s estimates can be compared to the teleseismic M_s of Eisenhauer (1976). These are plotted in Figure 2. To give some idea of the scale, the standard deviation of the mean residual between the two M_s measurements is less than 0.10 M_s units. This indicates that the Airy phase measurements from these two stations give M_s measurements nearly the same as the teleseismic M_s from many stations compiled by Eisenhauer.

The data plotted in Figure 1 indicate that the slope of M_s versus log W is not much different from unity. Then one way to indicate the relative coupling in different areas is to compute M_s - log W for each event. The mean values of this quantity are shown in Figure 3 for each population.

We see that the M_s and Airy phase data agree that the M_s coupling is about 0.1 - 0.3 units higher for events below the water table at Pahute Mesa than comparable events at Yucca Flat. There are only a few events above the water table at Pahute Mesa, but they clearly couple more weakly into M_s than those in saturated materials. The lowest M_s events are those in dry tuffs at Yucca Flat. The scatter is quite large for these low yield events, but M_s - log W is 0.57 lower, on the average, than for the saturated events at Yucca Flat. This difference is much larger than the standard

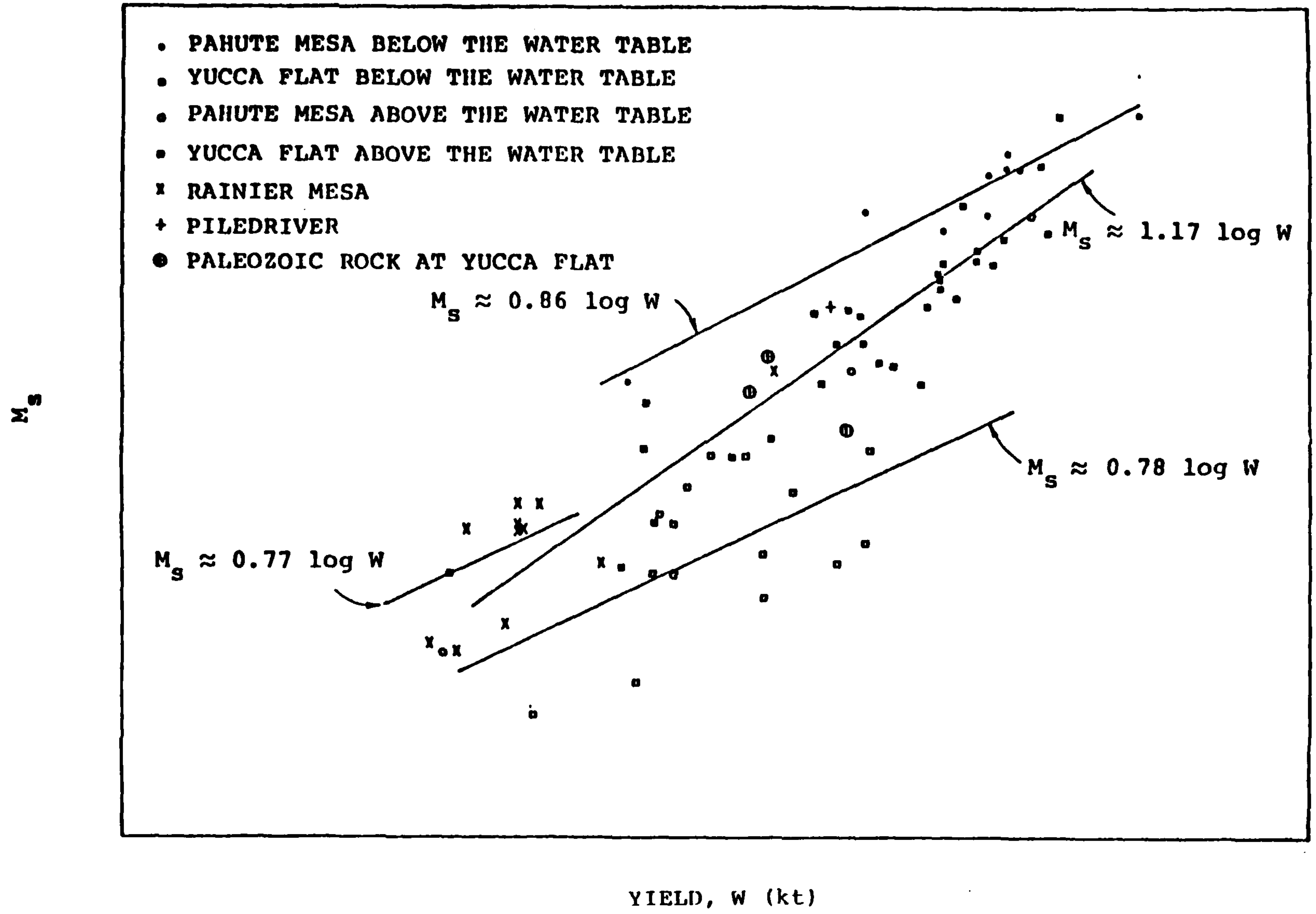


Figure 1. The M_s values from the WWSSN stations ALQ and TUC are plotted versus explosion yield. The explosions are divided into different populations according to the gross properties of the emplacement media.

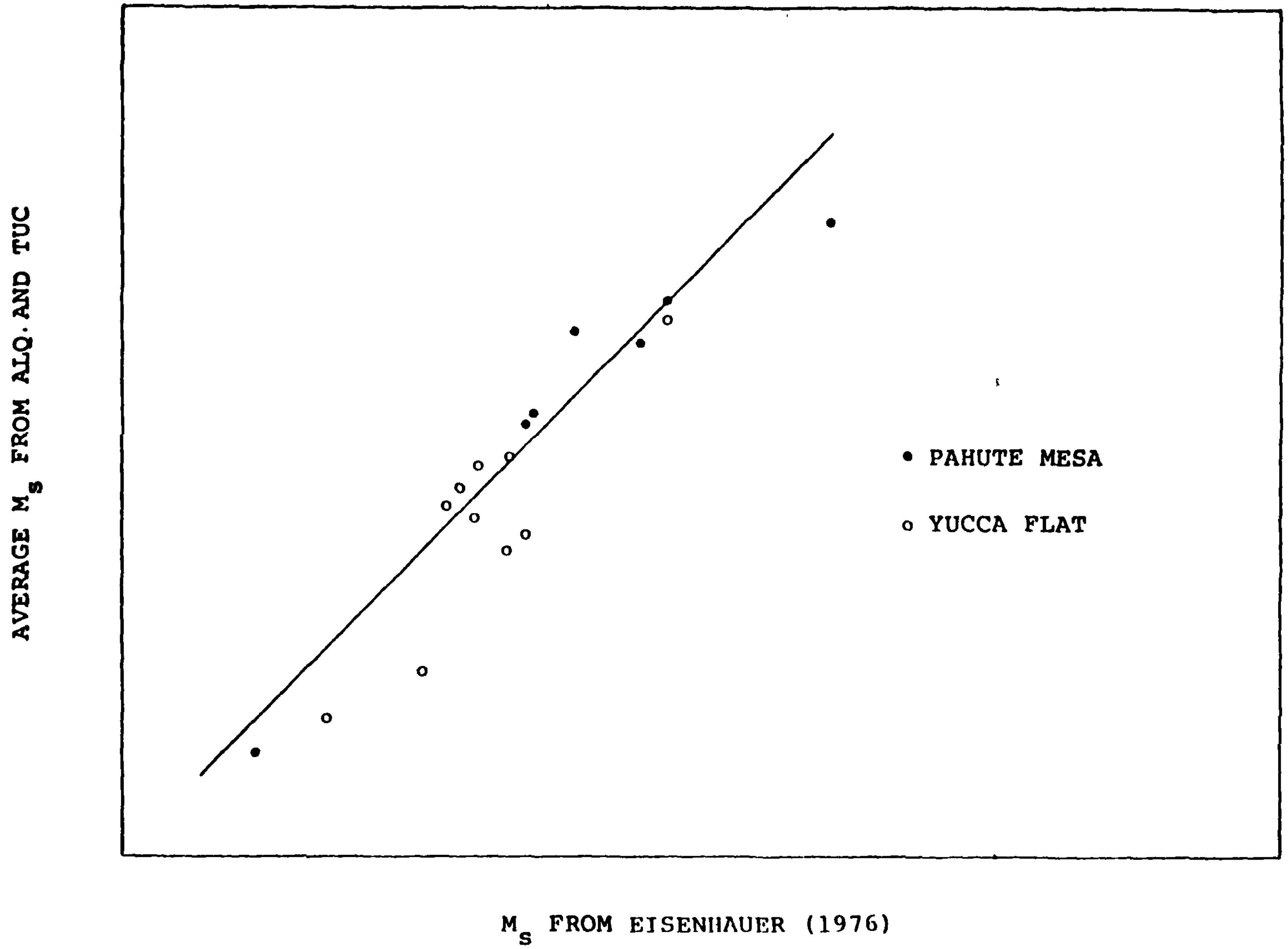


Figure 2. The M_s estimates from ALQ and TUC are compared to those from Eisenhauer (1976) for eighteen common events. A line of unit slope is shown for convenience. The standard deviation of the mean residual between these two M_s measures is 0.10 M_s units.

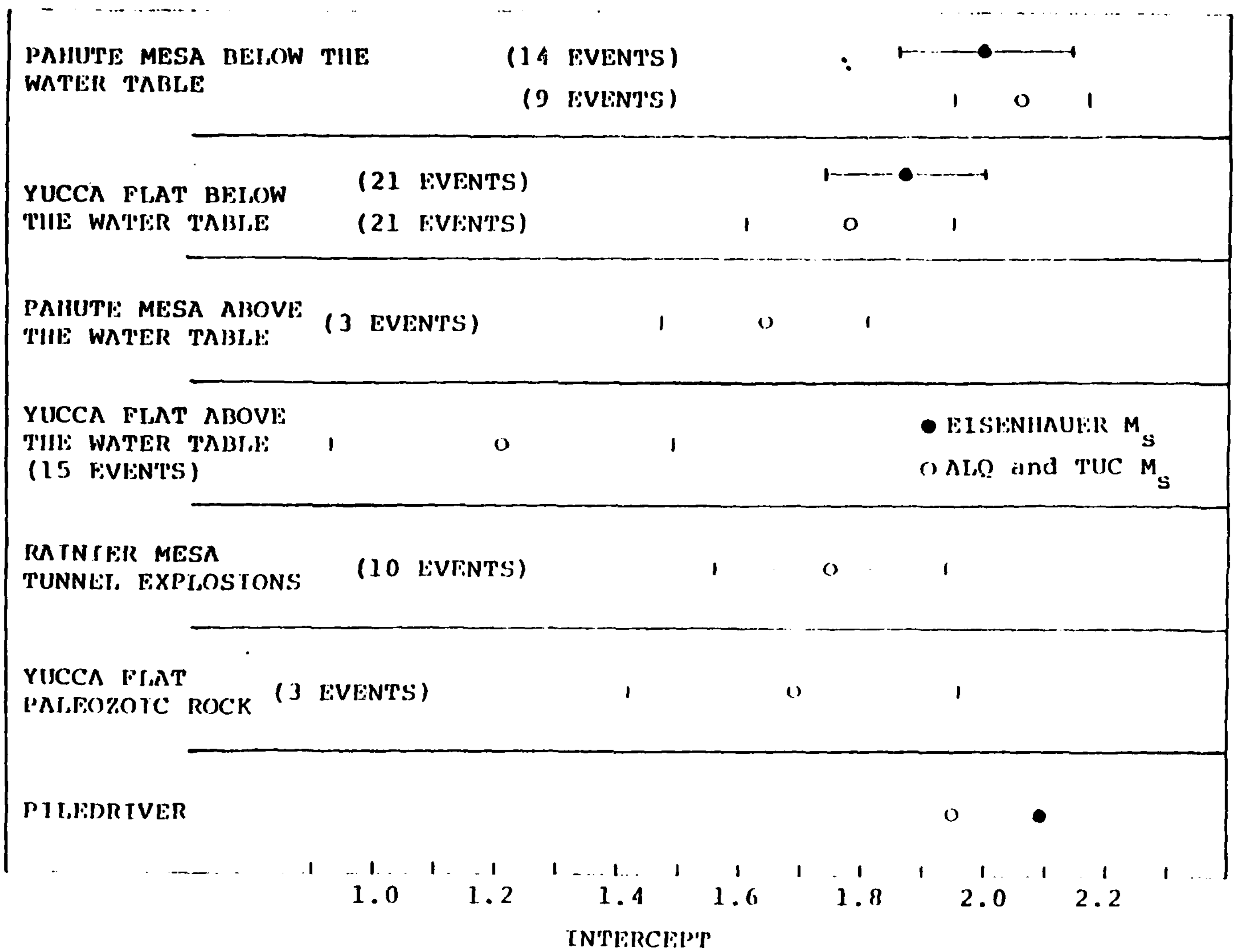


Figure 3. The mean values of M_S-log W for events in a limited yield range in each of the testing areas are depicted by a bar graph. For each data set the lines represent the mean $\pm$ one standard deviation.

deviation of the data. The saturated tuff explosions in the tunnel beds at Rainier Mesa seem to couple about the same as the saturated tuff events identified as being in Paleozoic rock at Yucca Flat. Since these events were detonated close to the tuff-Paleozoic interface, the identification of these events as being in Paleozoic rock is somewhat ambiguous and we are probably not seeing the true differences between events in the two source media. Finally, the PILEDRIVER event seems to couple like the highest coupling population, the Pahute Mesa events below the static water table.

An Explanation of Relative Coupling Differences

We have shown that there is a strong dependence of surface wave amplitude on the gross properties of the source materials. Particularly noticeable is the difference between saturated and unsaturated materials. Is this consistent with theory?

Under Topic 13 we discuss in some detail our synthetic seismogram studies of the ALQ and TUC data. Bache, Rodi and Harkrider (1978) and Bache, Rodi and Mason (1978), found crustal models that lead to synthetic seismograms in close agreement with the observations. The synthetic and observed seismograms were then compared to determine the apparent source amplitude. There is considerable discussion in those reports of higher order source effects (e.g., spallation, tectonic release), but a good first order theory is to assume the source can be represented by a reduced displacement potential that is flat at long periods.

During this study the dependence of Rayleigh wave amplitude on the source properties was carefully examined. As summarized under Topic 13, we concluded that the amplitude is

proportional to $\mu\Psi_\infty$ for similar materials like the Yucca Flat and Pahute Mesa saturated tuffs. When the local materials are much different, as for PILEDRIVER, the dependence on μ is much weaker and is frequency dependent.

The events carefully examined in that study were PILEDRIVER and a number of saturated tuff events at Yucca Flat and Pahute Mesa. The inferred Ψ_∞ values for the tuff events are shown in Figure 4, scaled to a constant yield. A correction for depth has been made and the yields of these events are in a narrow range from 40 to 200 kt.

How do the inferred Ψ_∞ values compare to independent estimates for the long period source level? This comparison is given in Table 1. Our inferred values appear to be within acceptable limits.

Unfortunately, we have not looked closely at the theoretical explanation of the relative surface wave amplitudes in other classes in Figure 3. The events above the water table have generally lower μ and lower Ψ_∞ than those below the water table. Therefore, the explanation for the relative coupling in these classes is as expected. The Rainier Mesa events have low μ but large Ψ_∞ (Bache, et al, 1975). Of course, the crustal structures are different and the scaling may not be simply $\mu\ \Psi_\infty$. The theoretical tools are available to look at this with minor effort.

Surface Waves from Finite Difference Source Calculations

In several of the topic summaries prepared for this assessment, I have pointed out the need for a more detailed understanding of the nonlinear coupling of the explosion energy into seismic waves. One-dimensional calculations have added a great deal to our understanding of the gross effects of source region material properties (e.g., air-filled

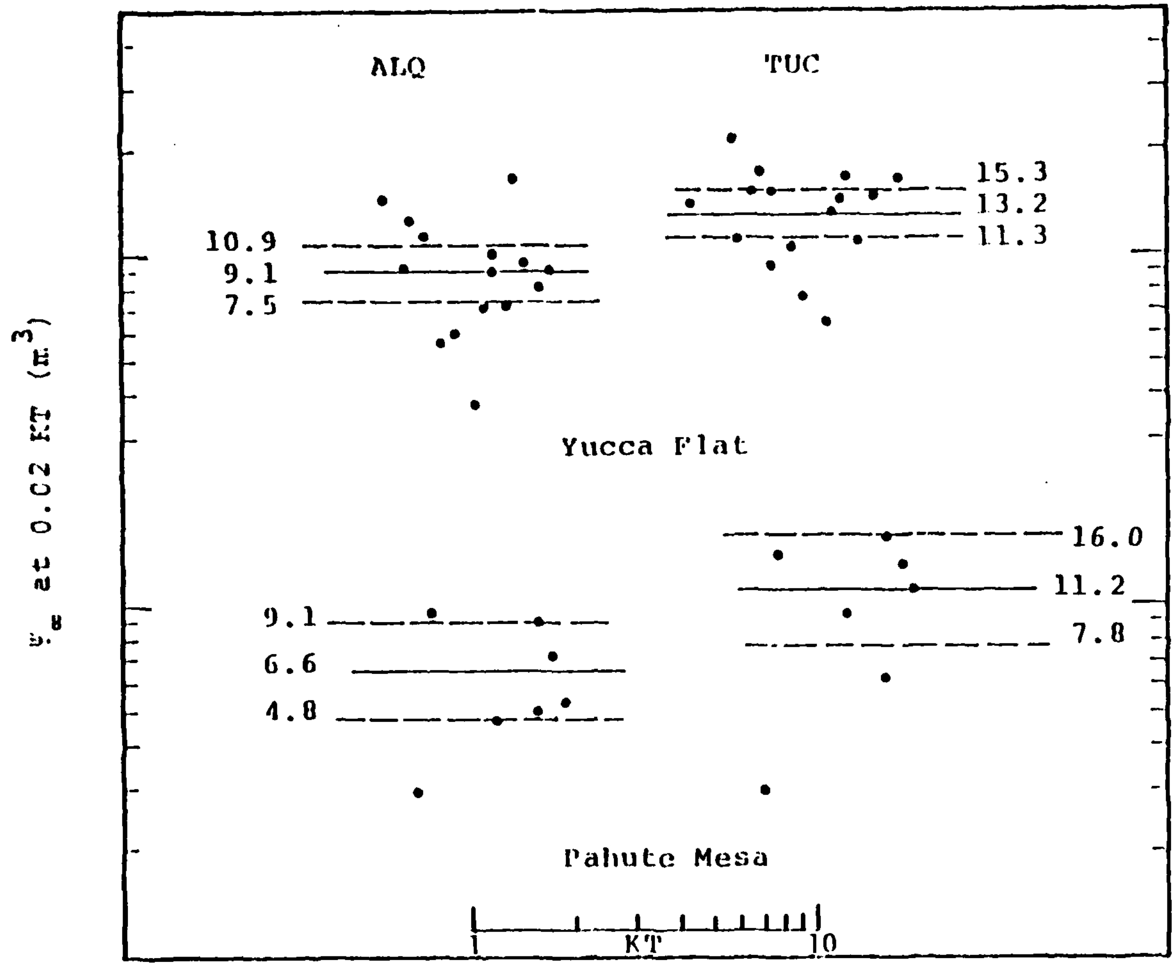


Figure 4. The Ψ_∞ values (at 0.02 KT) are plotted versus explosion yield for the Yucca Flat and Pahute Mesa events. The plot is log-log and the mean and 95 percent confidence limits on the mean are shown. One decade of yield is shown for reference.

TABLE 1

COMPARISON OF THEORETICAL AND OBSERVED Ψ_∞ VALUES TO THOSE INFERRED FROM THE ALQ AND TUC RAYLEIGH WAVES. ALL Ψ_∞ VALUES ARE SCALED TO 0.02 KT.

	Yucca Flat	Pahute Mesa	PILEDRIVER
ALQ Inferred	9.1 7.5 - 10.9 n = 17	6.9 4.8 - 9.1 n = 6	3.9
TUC Inferred	13.2 11.3 - 15.3 n = 18	11.2 7.8 - 16.0 n = 6	6.2
Inferred from moment values of Aki, *et al.*, 1974	16.6 13.2 - 20.7 n = 8	16.2 12.5 - 21.0 n = 4	3.75†
Inferred from cavity radius (Murphy, 1974)	15.1 13.8 - 16.6 n = 8	12.2 9.6 - 15.3 n = 4	-
Theoretical Values from S^3 Finite Difference Calculations	4.8, 6.8, 10.3**	4.2-7.2**	9.2
Observed Haskell (1967)	20.5*	-	10.0
RDP from PILEDRIVER Records	-	-	5.8, 16.7

*Haskell (1967) lists this value as questionable.

†Based on a yield of 56 KT.

**Range of values for expected range of material properties.

porosity, plastic yielding, etc.). The results of these one-dimensional calculations are easily studied in terms of the radiated seismic waves because they are easily represented by an equivalent elastic source, the reduced displacement potential.

Many interesting and potentially important source effects are two- or three-dimensional. These include the depth-dependence of overburden pressure and rock properties, nonlinear interaction with the free surface and spallation, complex geologic profiles and tectonic strain release. Techniques for detailed modeling of the nonlinear physical processes are available and improving. However, interpretation of the results has been difficult because of the lack of a satisfactory technique for propagating the computed elastic waves to distances of interest.

A straightforward technique for analytically continuing the results of finite difference source calculations has been developed at S^3. The development and its implementation for computing the surface waves from axisymmetric source calculations is described in a report by Rimer, et al. (1979). In that report we described the application of the technique to several test problems for which the exact Rayleigh wave solution is known.

The technique is simply to program the elas[illegible]amic representation theorem (Burridge and Knopoff, 19[illegible] [illegible]en no body forces are present this may be written

$$u_i^F = -\int_{S_M} \left[G_j^i * T_j^M - S_{jk}^i * u_j^M \, n_k \right] dA \, ,$$

$u_i^F(\underset{\sim}{x}_F, t)$: Components of the displacements at $\underset{\sim}{x}_F$ outside the source region;

$G_j^i(\underset{\sim}{x}_M, \underset{\sim}{x}_F, t)$: Displacement in the direction j at x_M due to a unit impulsive force at $\underset{\sim}{x}_F$ in the direction i;

$S_{jk}^i(\underset{\sim}{x}_M, \underset{\sim}{x}_F, t)$: The components of the stress tensor at x_M due to a unit impulsive force at $\underset{\sim}{x}_F$ in direction $\underset{\sim}{i}$;

$T_j^M(\underset{\sim}{x}_M, t)$: Components of the monitored traction vector on the monitoring surface;

u_j^M : Components of the monitored displacement vector on the monitoring surface;

n_k : Components of the normal to the monitoring surface, with the positive sense being for vector pointing <u>away from the source</u>;

S_M : Area of the monitoring surface.

The "*" indicates convolution.

The algebra for implementing this method is formidable and the programming and numerics are not easy. One of the test problems used to demonstrate the successful implementation of the technique is shown in Figure 5. A fully elastic explosion calculation was done using the SWIS finite element program. The stresses and displacements were calculated at the indicated stations. The Rayleigh pulse for the exact solution and the analytically continued solution with all (twenty) and half the monitoring stations is plotted in Figure 6. The spectral ratios with respect to the exact solution are shown in Figure 7. The long period solution turns out to be the small difference between two large cancelling contributions to the quadratures. Hence the importance of having a dense spatial sampling at long periods.

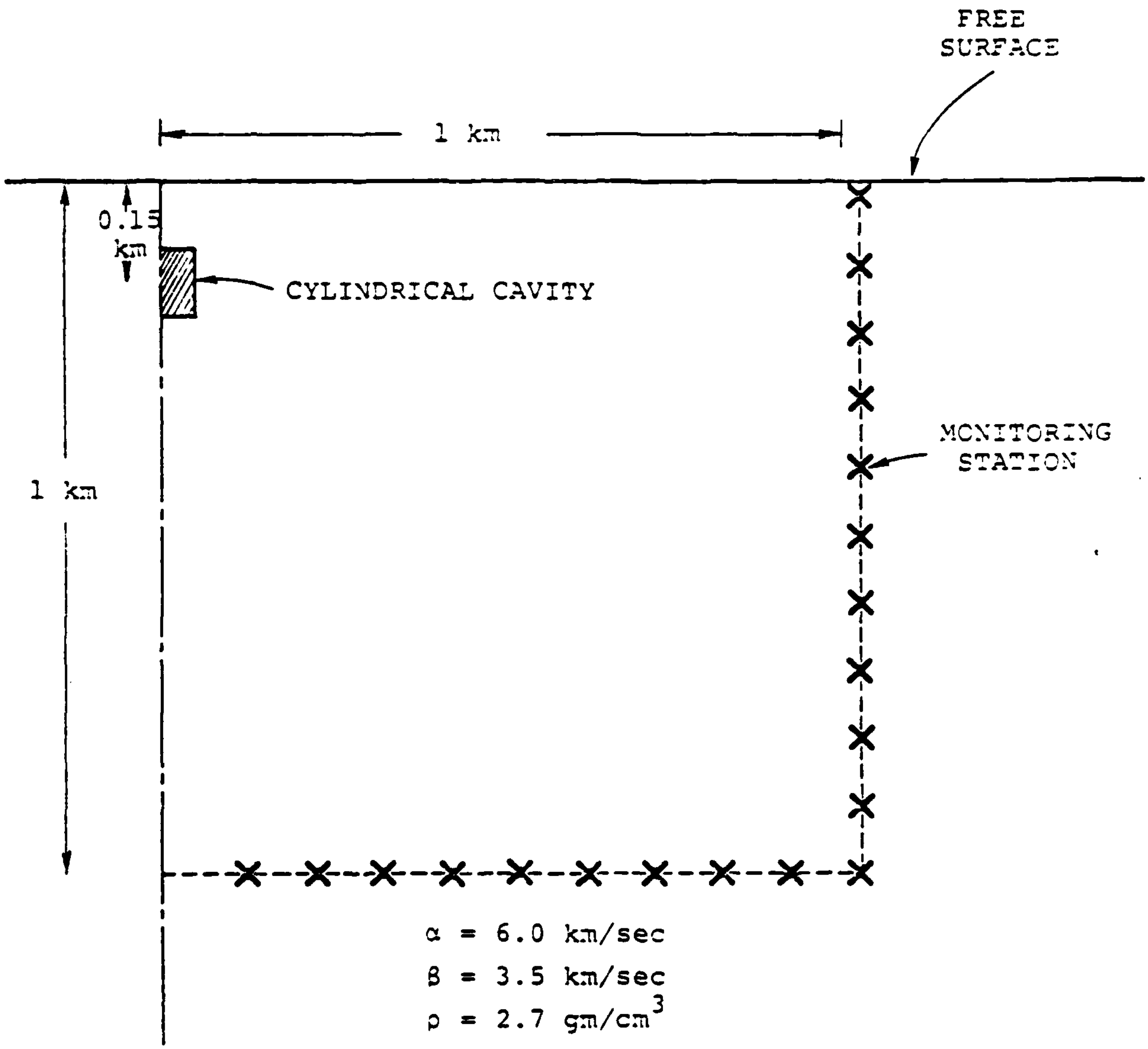


Figure 5. Schematic view of the axisymmetric grid for the SWIS calculation. The twenty monitoring solutions are indicated.

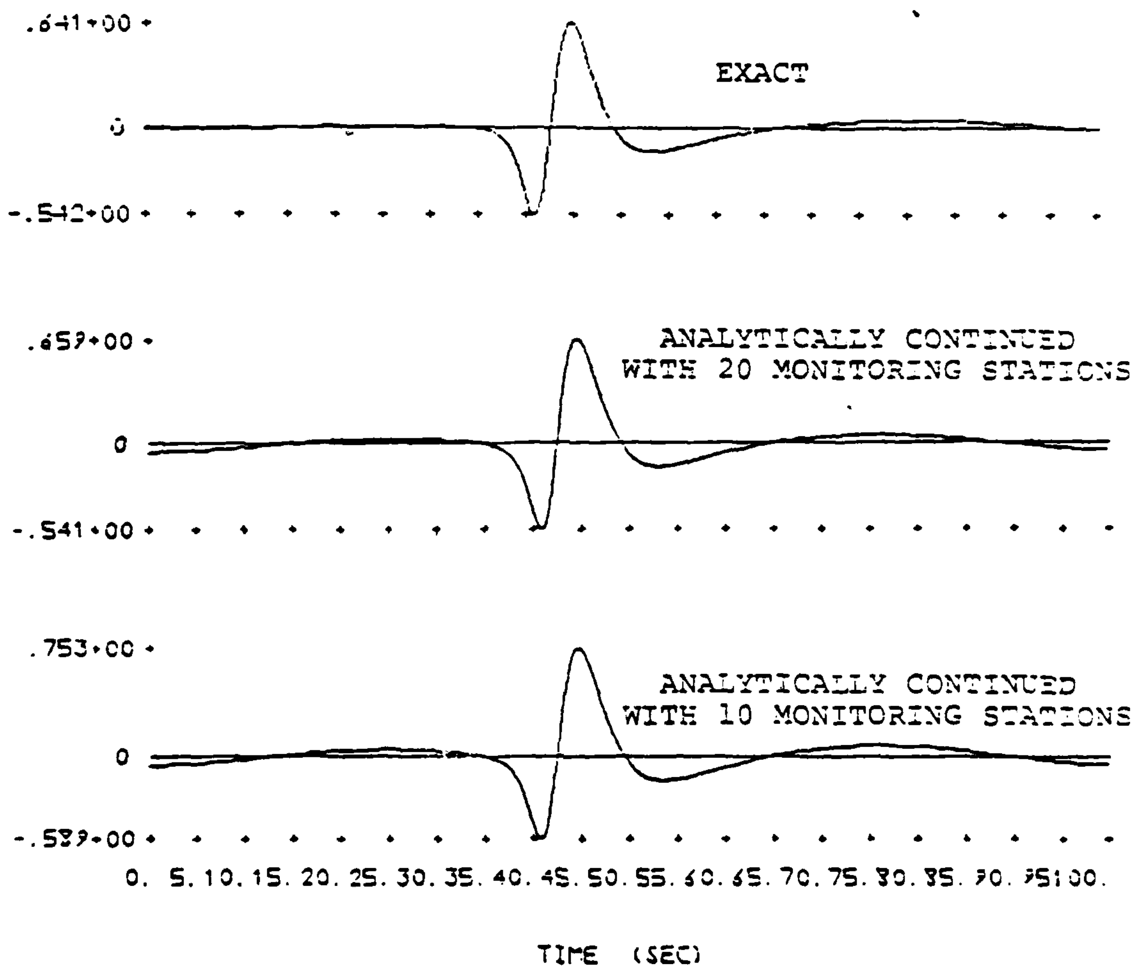


Figure 6. Comparison of exact and analytically continued halfspace Rayleigh waves after propagation to 1,000 km and filtering by the WWSSN 15-100 instrument response.

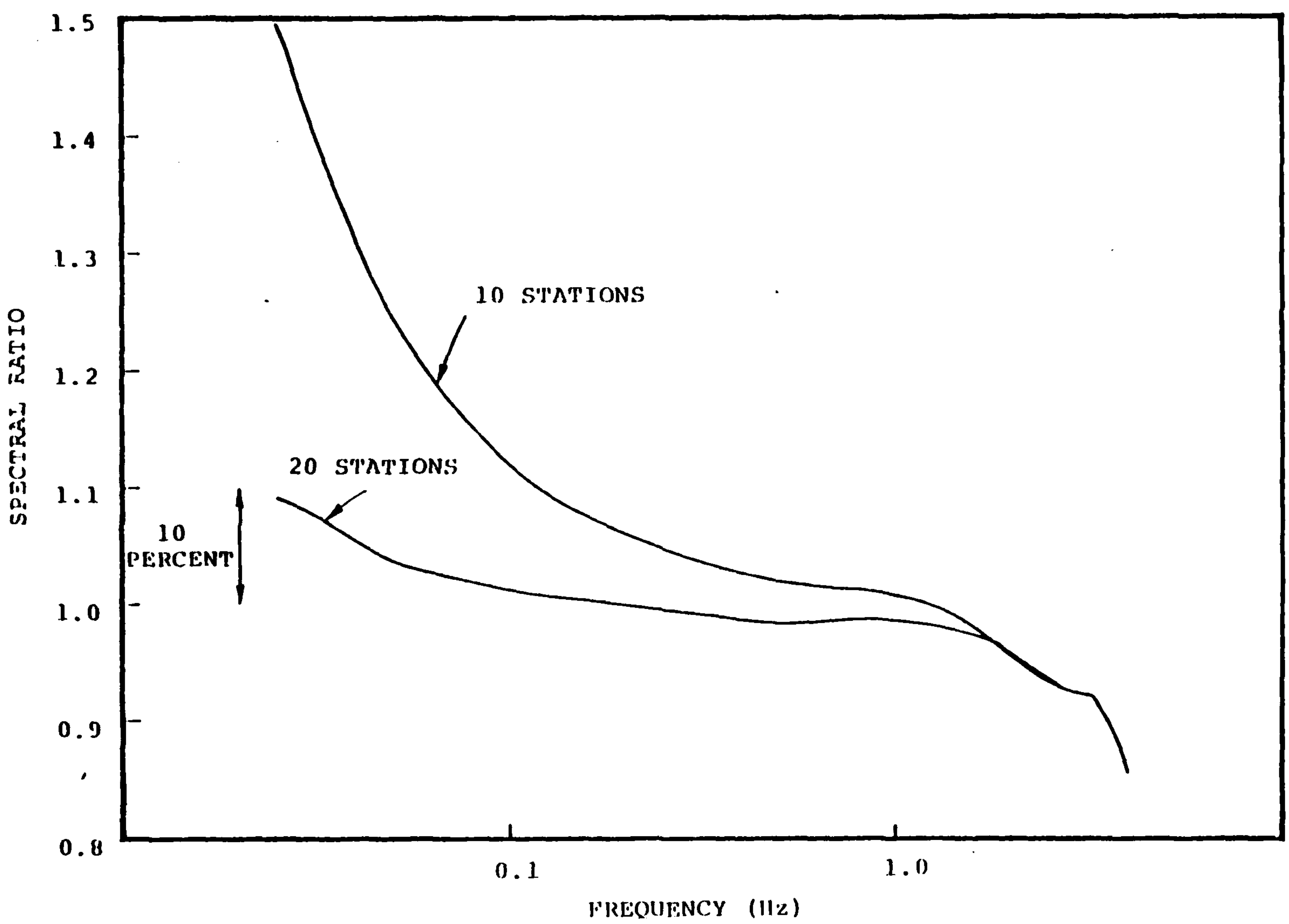


Figure 7. Spectral ratio: the analytically continued numerical solution divided by the exact solution for the halfspace Rayleigh wave.

Surface Waves from Detailed Models of Granite Explosions

The technique outlined and its analogous implementation for body waves are being used to study two-dimensional finite difference calculations of explosions in granite and salt by S^3 and granite and sandstone by Applied Theory, Inc. The study is not complete and no firm conclusions can yet be drawn. However, we can indicate the kinds of results we are getting.

The S^3 granite calculations are meant to model PILEDRIVER and three similar events. The PILEDRIVER calculation was quite successful in matching a wide variety of near field ground motion data. For example, some measured and calculated time histories are plotted in Figures 8 through 11.

Synthetic seismograms computed with an RDP source were compared to observations of PILEDRIVER by Bache, Rodi and Harkrider (1978). The waveform comparison is shown in Figure 12. Also shown is the comparison of the observations to synthetic seismograms computed by analytically continuing the two-dimensional finite difference source calculation. At these periods the two-dimensional effects have little influence on the waveform. This is not unexpected. Bache, Rodi and Mason (1978) found the ALQ and TUC waveforms to be quite insensitive to source perturbations such as the addition of a spall impulse source and/or reduction of the amplitude of the upgoing waves from the source. However, the seismogram amplitude was found to be sensitive, especially to the attenuation of upgoing waves.

The predicted amplitude is about twice as large as that observed. This indicates that either our source is too large or that our path corrections are wrong. We would like to think that we have accounted for the path with considerable accuracy.

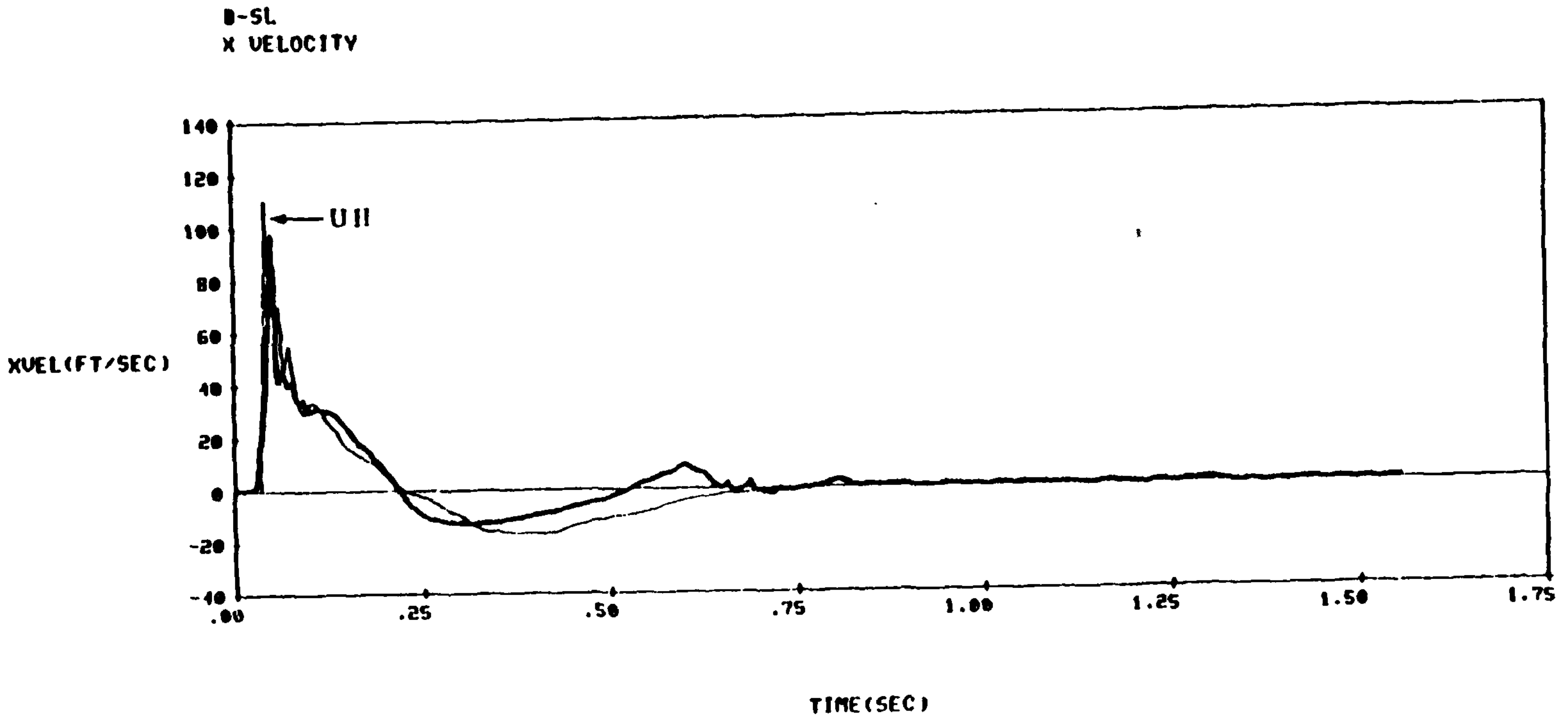


Figure 8. Measured and calculated horizontal velocities at shot level Perret station B-SL (range = 204 m).

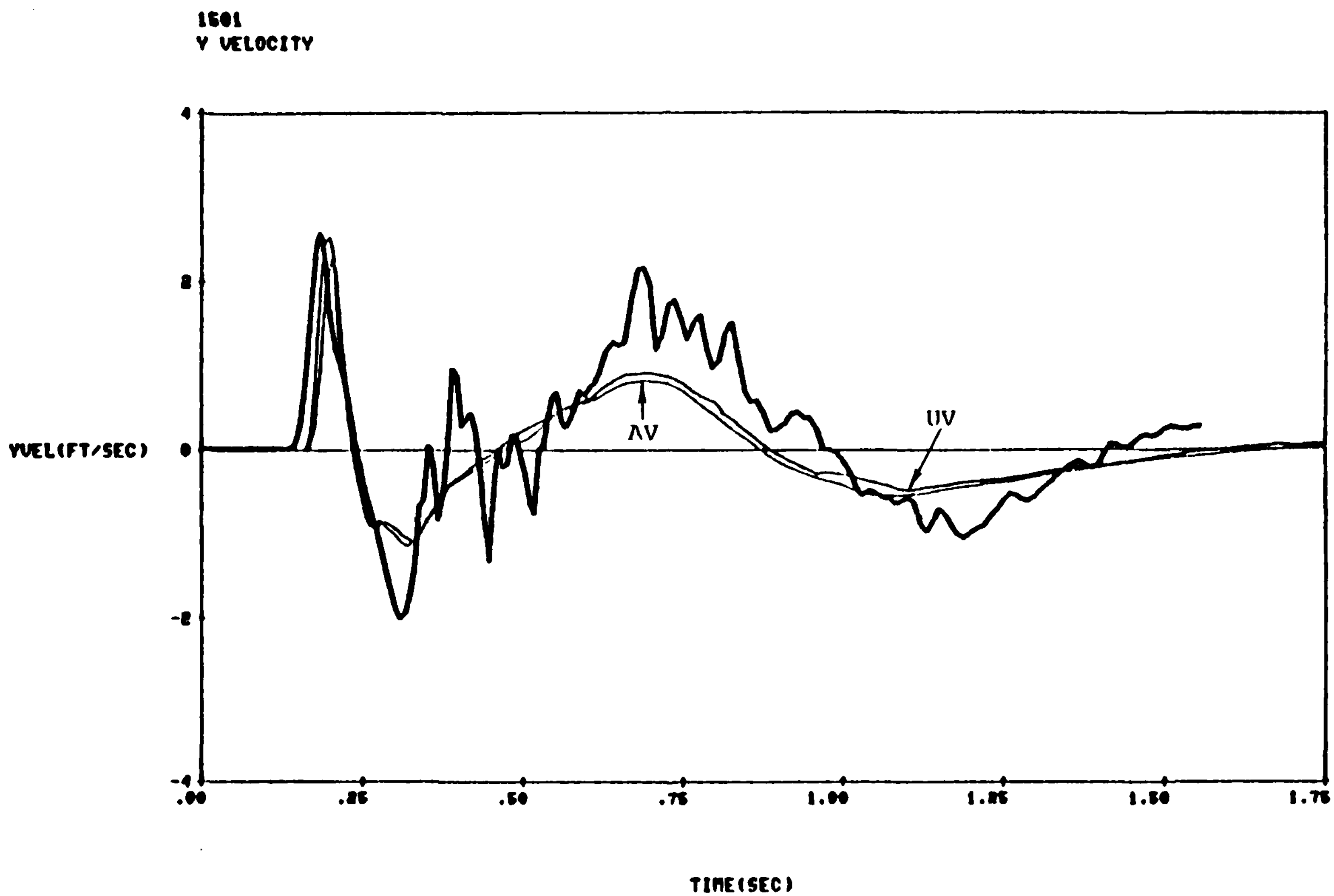


Figure 9. Measured and calculated vertical velocities at station 1501, 215 m above shot level (horizontal range = 863 m).

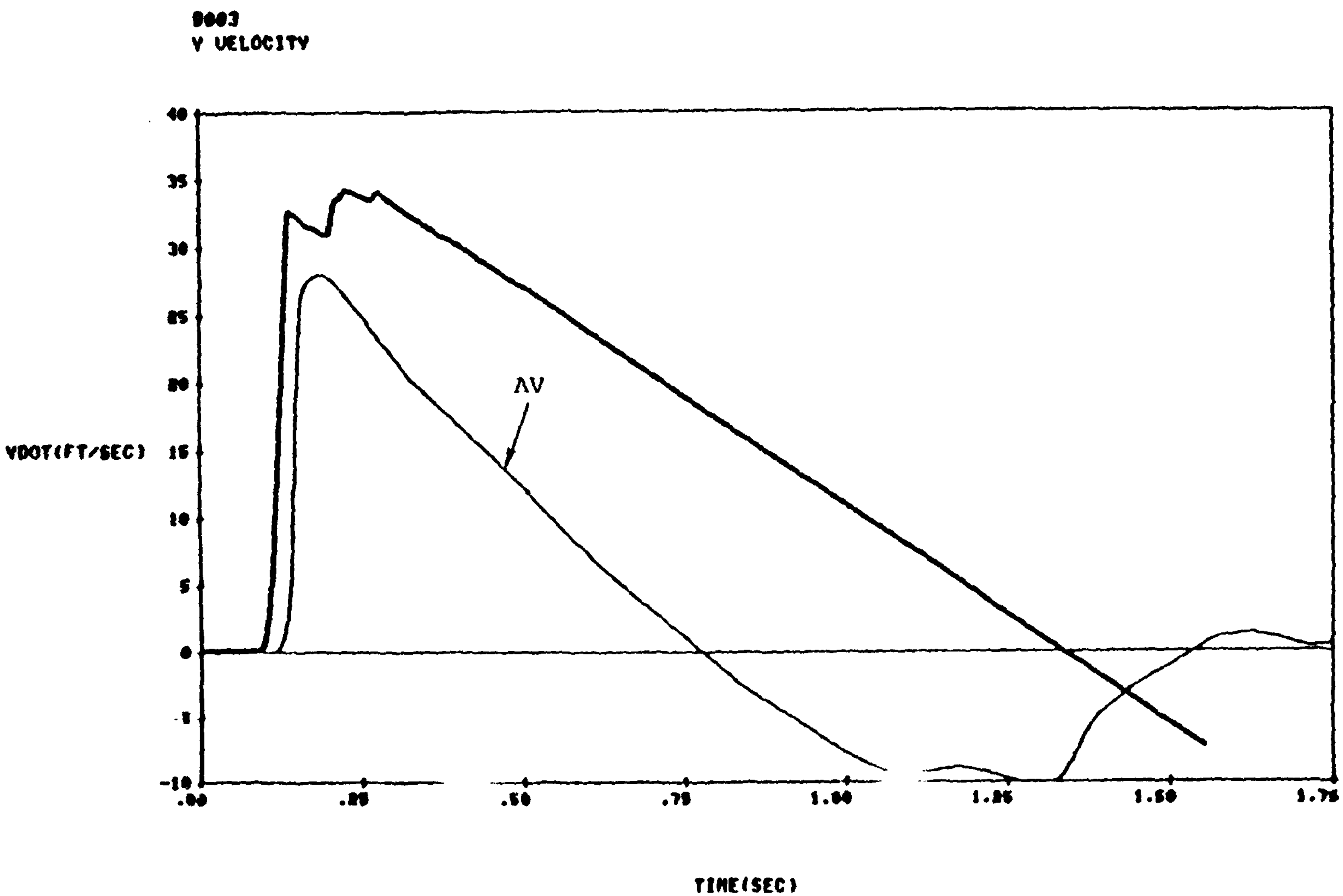


Figure 10. Measured and calculated vertical velocities at free surface station 9003 (horizontal range = 110 m).

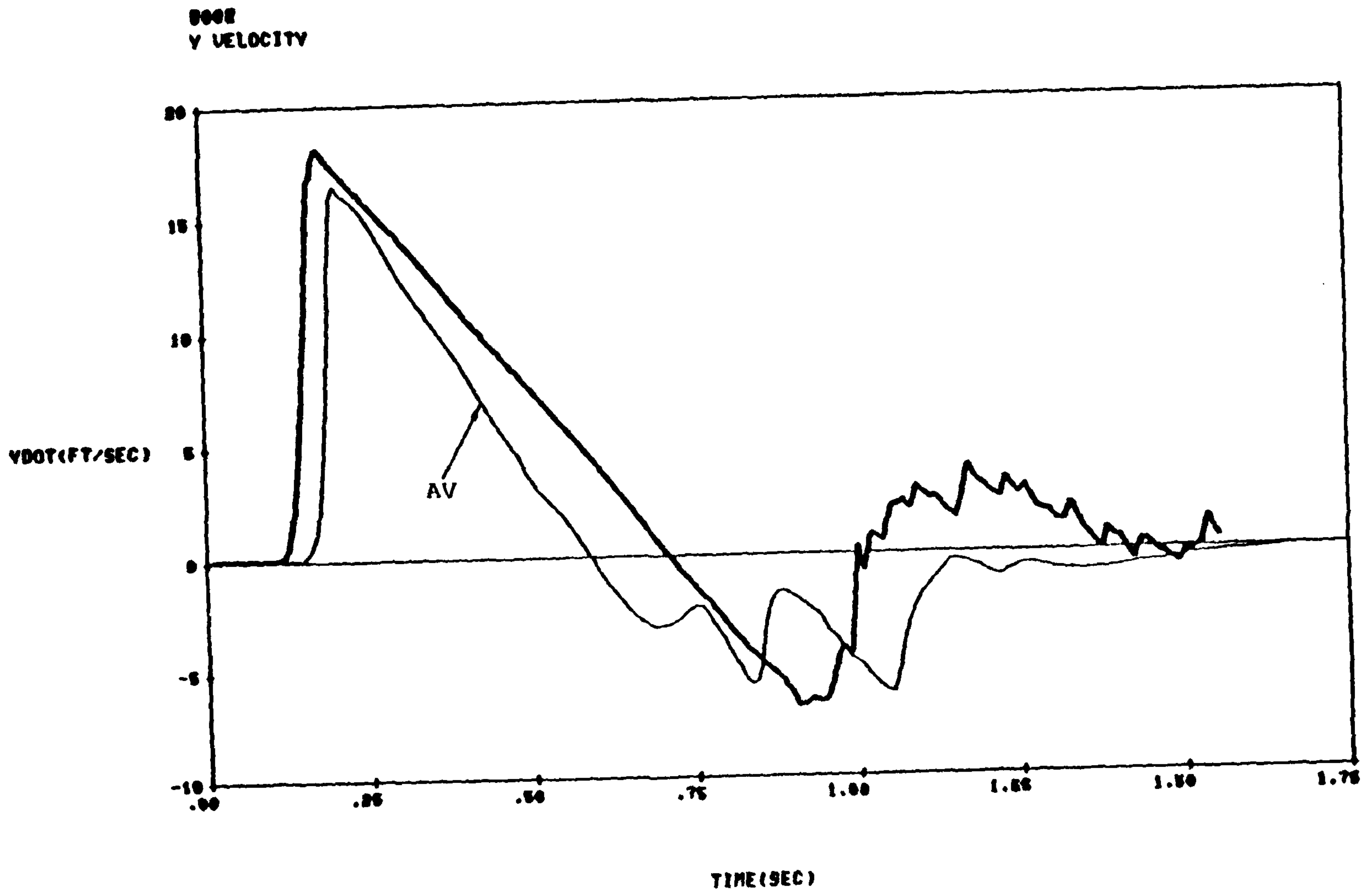


Figure 11. Measured and calculated vertical velocities at free surface station 9008 (horizontal range = 432 m).

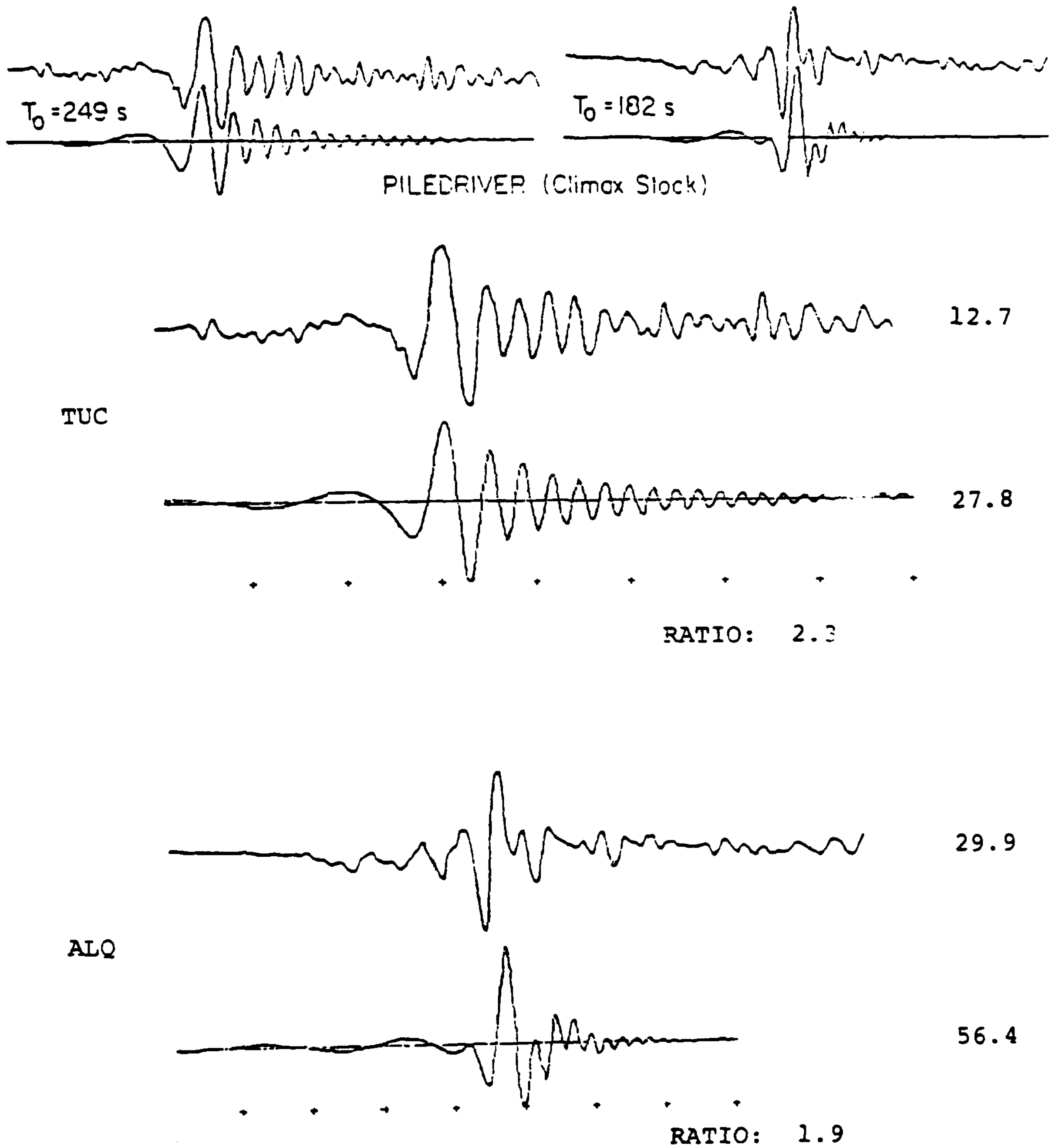


Figure 12. Observations of PILEDRIVER from ALQ and TUC are compared to synthetic seismograms. The synthetics are on the bottom for each pair. The two sets at the top of the page are from Bache, et al. (1978) and were computed with an RDP source. For the others the synthetics were computed by analytically continuing the two-dimensional PILEDRIVER calculation. The peak-to-peak amplitude in microns is listed with each record.

Is the two-dimensional source too large at long periods? It could be; the comparison with the data in Figures 8 through 11 emphasizes the agreement at high frequencies. However, more important is likely to be the absence of the tectonic release contribution. The moment tensor fit to the PILEDRIVER surface waves by Rivers and von Seggern (1978), discussed under Topic 11, suggests that the double-couple causes the radiated waves at these azimuths to be smaller than otherwise by a factor of 1.5 to 2.0. Adding this double-couple would bring the observed and synthetic seismograms into close agreement.

Another way to look at the two-dimensional calculation is in terms of the RDP one would infer from the surface wave seismogram if the path amplification were known. For our synthetic seismograms we know the path perfectly and can easily determine this "equivalent source" $\hat{\Psi}_e$. This is the RDP source that gives the same Rayleigh wave as the two-dimensional calculation. The $\hat{\Psi}_e$ for the PILEDRIVER calculation is shown in Figure 13 where it is compared to two other estimates of the PILEDRIVER RDP. One is the Mueller/Murphy (1971) source for granite. The other is from a spherically symmetric calculation of PILEDRIVER with the same constitutive models for the source region granite. Comparison of the $\hat{\Psi}_e$ with the latter directly displays the two-dimensional effects.

Note that the long period level of $\hat{\Psi}_e$ is about 11 m^3, scaled to 0.02 kt. The Ψ_∞ values from the Bache, Rodi and Mason (1978) study given in Table 1 are 3.9 m^3 at ALQ and 6.2 m^3 at TUC. These are the Ψ_∞ values required to bring the synthetic and observed amplitudes into agreement and are consistent with the fact that the two-dimensional study gives Rayleigh wave amplitudes that are about twice as large as those observed.

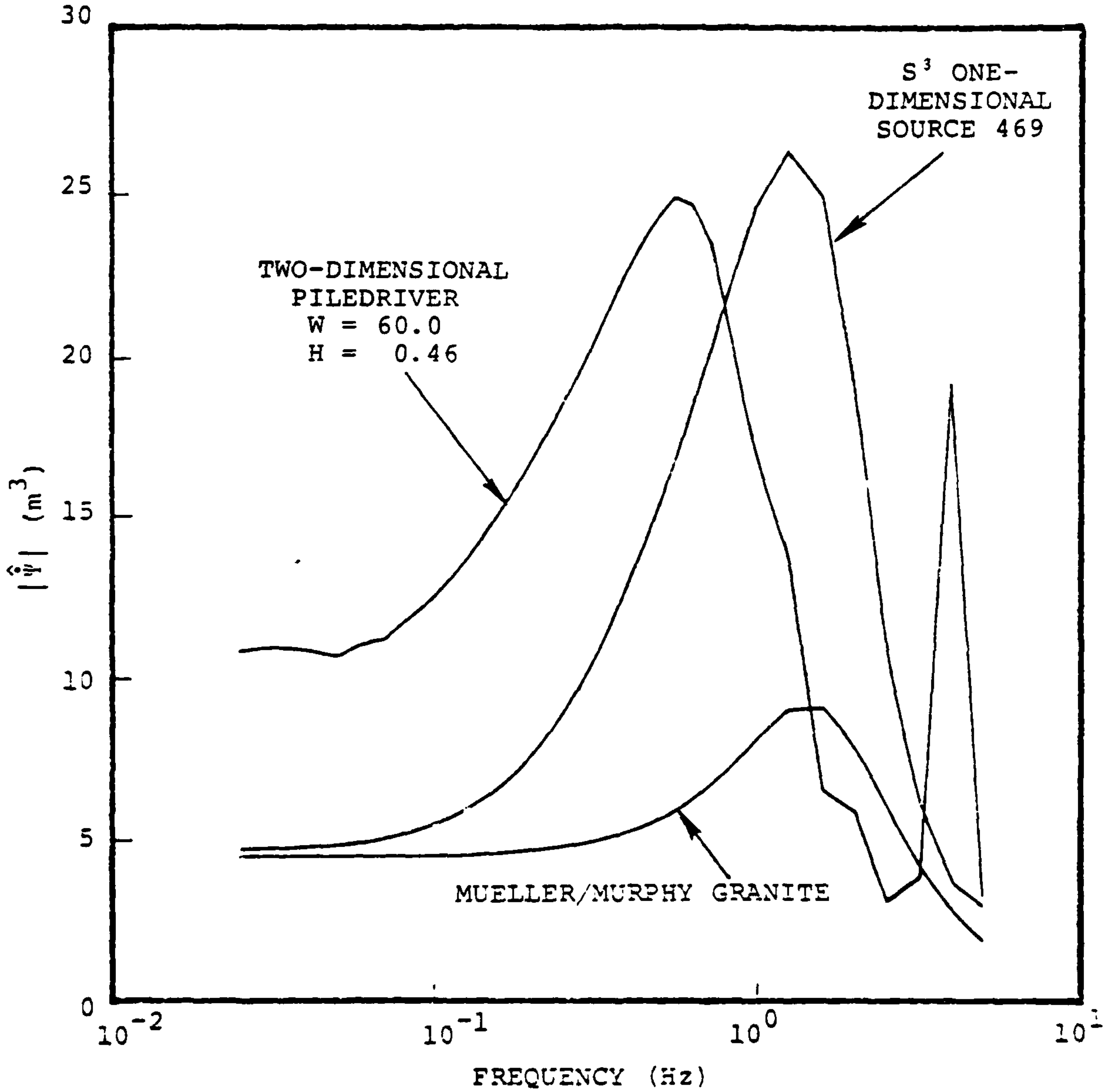


Figure 13. The "equivalent RDP," $\hat{\Psi}_e$, for the two-dimensional PILEDRIVER calculation is compared to two other estimates for the RDP.

Conclusions

Surface wave amplitude data show that the amplitude data separate into classes dependent on the gross properties of the local source material. The difference between dry and saturated materials is particularly striking. The differences are essentially the same whether one uses a multistation M_s to indicate amplitude or uses Airy phase amplitudes from stations like ALQ and TUC. These stations are useful because they record more of the low coupling events.

The amplitude dependence on source properties seems consistent with our theoretical models for source coupling into Rayleigh waves. That is, estimates for the dependence of Ψ_∞ on the source properties and the dependence of Rayleigh waves excitation on the crustal layering are consistent with the observations. A bit more work is required to quantify this dependence for the Rainier Mesa tuffs and the unsaturated tuffs at Pahute Mesa and Yucca Flat, but it seems to be qualitatively about as expected.

The RDP models explain the amplitude variations between populations of events, but there are many observed phenomena that are not consistent with this sample model. These include tectonic release and phenomena, such as spallation, related to the depth dependence of material properties and overburden pressure. More detailed source calculations including the physics associated with these complex phenomena are required to develop a better understanding of their influence on Rayleigh wave excitation.

Finite difference calculations for modeling explosions in two, or even three dimensions, have been under development for many years. Results from these calculations have been compared to near-field, high frequency data and reasonable results can be obtained. For seismic yield determination, the usefulness of these detailed models has been questionable

because the coupling into far-field seismic waves could not be estimated very accurately. However, a technique for analytically continuing the results of finite difference source calculations to compute synthetic seismograms in realistic earth models has recently been implemented at S^3.

The use of multi-dimensional finite difference source calculations for studying seismic coupling is beginning. In this summary we show one particular example, a two-dimensional calculation of PILEDRIVER. This illustrates the technique and shows the kinds of results we can expect. There is good cause to be optimistic that these complex calculations can fill in many of the remaining open areas in our understanding of the seismic coupling of explosions.

REFERENCES

Bache, T. C., J. T. Cherry, N. Rimer, J. M. Savino, T. R. Blake, T. G. Barker and D. G. Lambert (1975), "An Explanation of the Relative Amplitudes Generated by Explosions in Different Test Areas at NTS," Systems, Science and Software Final Report to the Defense Nuclear Agency, DNA 3958F, October.

Bache, T. C., P. L. Goupillaud and B. F. Mason (1977), "Seismic Studies for Improved Yield Determination," Systems, Science and Software Quarterly Technical Report SSS-R-77-3345 submitted to AFTAC/VSC, July, 65 pages.

Bache, T. C., W. L. Rodi and D. G. Harkrider (1978), "Crustal Structures Inferred from Rayleigh-Wave Signatures of NTS Explosions," BSSA, 68, pp. 1399-1413.

Bache, T. C., W. L. Rodi and B. F. Mason (1978), "Source Amplitudes of NTS Explosions Inferred from Rayleigh, Waves at Albuquerque and Tucson," Systems, Science and Software Report SSS-R-78-3690 submitted to ARPA/VSC, June, 91 pages.

Burridge, R. and L. Knopoff (1964), "Body Force Equivalent for Seismic Dislocation," BSSA, 54, pp. 1875-1888.

Eisenhauer, T. D. (1976), "M_s and m_b Estimates for USSR Underground Explosions," unpublished report.

Mueller, R. A. and J. R. Murphy (1971), "Seismic Characteristics of Underground Nuclear Detonations," BSSA, 61, pp. 1975.

Rimer, N., J. T. Cherry, S. M. Day, T. C. Bache, J. R. Murphy and A. Maewal (1979), "Two-Dimensional Calculation of PILEDRIVER, Analytic Continuation of Finite Difference Source Calculations, Analysis of Free-Field Data from MERLIN and Summary of Current Research," Systems, Science and Software Report SSS-R-79-4121 submitted to AFTAC/VSC, August, 158 pages.

Rivers, W. and D. H. von Seggern (1978), "Release of Tectonic Strain Energy by PILEDRIVER," unpublished Teledyne Geotech Memorandum, September.

von Seggern, and Gurski (1976), has measured Rayleigh wave amplitudes for all events large enough to be measured at MNNV during its operational life in the LRSM network. He has perfomred a multiple regression on the data including medium, site, and watertable effects. The results are much as expected with larger amplitudes below than above the water table, and in hard-rock than in allumvium. One perhaps surprising conclusion is that events in Pahute Mesa have 0.2 M_s lower amplitudes than otherwise equal events in Yucca Flats. This may reflect a difference in the Tuff at the two sites. Ramspott and Howard (1975) find a higher velocity for tuff at Pahute than at Yucca.

Ramspott, L. D., and N. W. Howard, (1975). Average properties of nuclear test areas and media at The Nevada Test Site UCRL-51948, Lawrence Livermore Laboratory, Livermore, California.

von Seggern, D. W., J. Gurski, (1976). M_s or yield of underground nuclear explosions at The Nevada Test Site, TR-76-11, Teledyne Geotech, Alexandria, Virginia 22314.

California Institute of Technology, Pasadena

Effect of Source Region Properties on Surface Wave Generation

David G. Harkrider

In order to investigate the effects of nonlinear regions above and below the source such as spall and cracking, Harkrider and Bache (1979) have developed a wave field separation that can be used in most computer programs based on residue theory, steepest descent approximation or wave number integration in terms of the up and downgoing waves from the source. This technique allows one to model the resultant source wave in the linear homogeneous regime by source radiation whose history, amplitude, or travel time differs above and below the source point.

When comparing Rayleigh wave generation for two source region models of NTS, Bache et al. (1977) found that the polarity of the Rayleigh waves generated by the downgoing radiation in the soft tuff over hard basement model were opposite from those from a source region in which the soft rock-hard rock contrast below the source is absent. If the Rayleigh waves from the upgoing source radiation or the source wave itself were degraded by surface or near surface irregularities or momentum traps then it would be possible to reverse the polarity of Rayleigh waves at all azimuths from explosions in the same general source area. This effect would be similar to that due to the superposition of the tectonic surface wave radiation from a 45° dipping pure thrust fault like mechanism.

In Bache et al. (1978), the source suppression technique was used to model spall events at NTS. The results were interesting if not conclusive,

in that the Piledriver source estimate is strongly influenced by two poorly understood effects: the superimposed double couple and possible suppression of the elastic response of the upgoing wave due to spall.

References

Bache, T. C., P. L. Goupilland, and B. F. Masson (1977), Studies for improved yield determination, Systems,Science and Software Quarterly Technical Report SSS-R-77-3345.

Bache, T. C., P. L. Goupilland, and B. F. Masson (1978), Source amplitudes of NTS explosions inferred from Rayleigh waves at Albuquerque and Tucson, Systems, Science and Software Topical Report SSS-R-78-3690.

Harkrider, David G. and Thomas C. Bache (1979),Rayleigh waves generated by the suppression of upgoing or downgoing explosion source waves, Annual Technical Report 1 April-30 September 1979, California Institute of Technology, AFOSR #F49620-77-C-0022.

TOPIC 10

SOURCE THEORY AND OBSERVATION FOR SURFACE WAVES

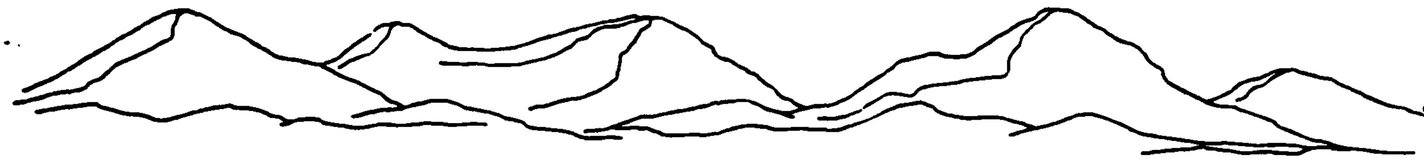
by

Lee Woodie

Pacific Sierra Research Corporation, Santa Monica, CA

During the years 1977-1979, we at PSR worked with Jack Trulio and Neil Perl of ATI to develop the most accurate state-of-the-art prediction algorithm for surface wave amplitudes now in use for underground explosions. The non-linear source model of ATI numerically simulates the rock motion from the blast volume out to a distance where linear elastic motion is expected, approximately 1 to 2 kilometers from the epicenter. From this range, a new linear elastic source is created involving a numerical grid of ring elements and which is suitable to act as a source term for surface wave Green's function propagators. Figures 1 and 2 on the following page show the numerical grid and the propagation scheme, respectively, which is fully described in PSR Note 233 [1]. Nearly the entire algorithm has been tested and verified against an ideal-Lamb's-problem-source [2], and so we feel very confident of the results. A few of the relevant details will now be discussed.

The explosion is assumed to take place in the top layer of a multi-layered half space, but far enough away from the first interface so that the rock motion is identical to a uniform half space explosion. The linearized source is composed of from 45 to 95 rings on which vertical and horizontal time dependent forces are allowed to act. These forces are suitable source terms for the point force Green's functions as presented by Harkrider [3]. At this point two simplifying assumptions are made concerning Harkrider's Green's functions: 1) the depth dependence in the source region is the same as for a uniform half space; 2) the frequency dependence is accessible through the (empirical) dispersion relation only. These two assumptions are accurate to within about 5 percent for fundamental mode Rayleigh waves and they allow us to bypass a full multilayered calculation. The dispersion relation presented by McEvilley [4] for North America was used, which shows no Airy phase near 20 sec periods. Finally,



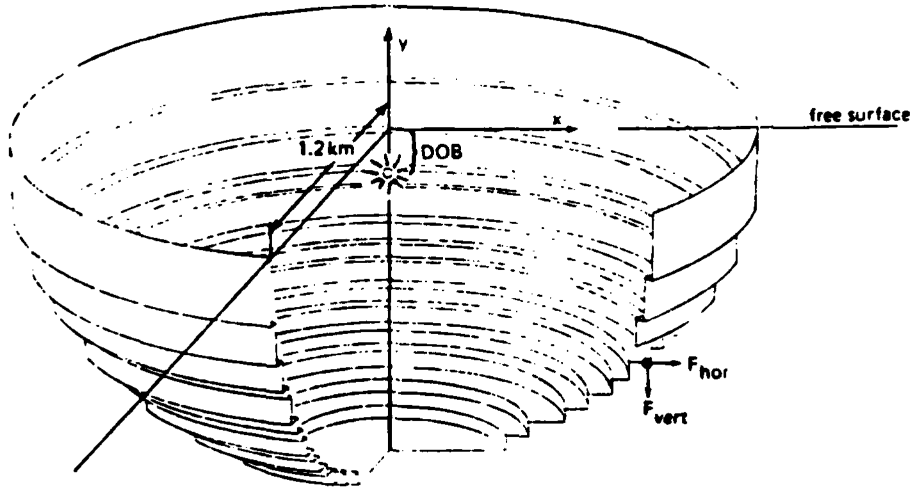


Fig. 1--Grid elements used to model magic-circle data of source volume for wet sandstone. Two force components are shown for one of the rings at one instant of time. DOB = depth of burial.

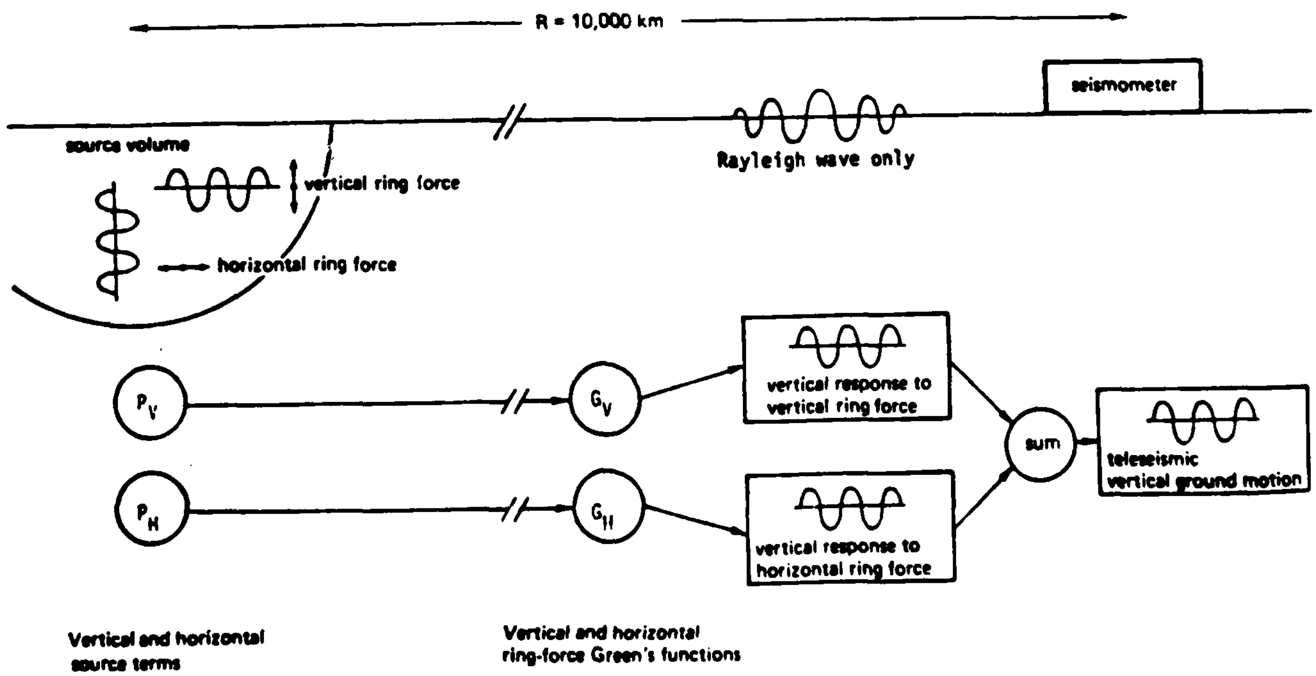


Fig. 2--Schematic outline of PSR M_s model showing source terms, propagators, and vertical response for one harmonic component and one ring force. Complete vertical response requires a summation over all rings and all harmonic components.

in order to obtain a better absolute value of M_s, a constant Q model is used where Q is independent of frequency and position. We use the value Q = 300. The complete model, as outlined in Fig. 2, was then used to compute maximum vertical teleseismic ground displacement, U_{max}, for a variety of depths in 3 different media, from which M_s was computed for 20-sec periods at a range of 10,000 km using a formula from Marshall & Basham [5]

$$M_s = 9.2 + \log\,(U_{max}\ [\mathrm{cm}]).$$

The modelled sources were computed over a depth range from 159 m to 1,000 m in simulated granite, wet sandstone, and "elastic" wet sandstone, all at yields of 150 KT. The finite element simulation of rock motion was carried out to about 2.2 sec, but a method of extrapolation to account for ejecta effects out to 20 sec periods was used. For the most part this extrapolation method insurred that the net vertical impulse was identically zero, since surface wave generation is very sensitive to non-zero vertical impulse. We found that M_s is not particularly sensitive to depth of burial and for any one of the three media $M_s \approx$ const ± 0.1 unit over the entire depth range. The results for 150 KT explosions were: granite M_s = 4.07, wet sandstone M_s = 3.68, "elastic" wet sandstone M_s = 3.46.

REFERENCES

1. Woodie, W. L., "Seismic Surface-Wave Propagation Model Using Surface-Wave Green's Function," PSR Note N233, November 1979.

2. Woodie, W. L., and S. Sanemitsu, "Surface Wave Test Case," PSR Note N295, August 1979.

3. Harkrider, D. G., "Surface Waves in Multilayered Elastic Media. I. Rayleigh and Love Waves from Buried Sources in a Multilayered Elastic Half-Space," *Bull. Seism. Soc. Am.*, Vol. 54, 1964, p. 627.

4. McEvilly, T. V., "Central U.S. Crust--Upper Mantle Structure from Love and Rayleigh Wave Phase Velocity Inversions," *Bull. Seism. Soc. Am.*, Vol. 54, 1964, p. 1997.

5. Marshall, P. D., P. W. Basham, "Discrimination Between Earthquakes and Underground Explosions Employing an Improved M_s Scale," *Geophys. J.R. Astro. Soc.*, Vol. 28, 1972, p. 431.

State of the Art
Tectonic Generation
Robert Blandford
Teledyne Geotech

Despite much excellent work by investigators in the field it seems to me that the true nature of the horizontally polarized shear waves emitted from explosions in earth materials is still a mystery. Futhermore it is not clear just what a suitable experiment would be to resolve the mystery. In this situation it seems to me that if ARPA would like to solve the problems the best course would be to first commission a comprehensive review of the literature, and then to give substantial unstructured grants of money to two or more interested investigators.

A fair amount of the literature was reviewed in the report by Blandford and Clark (1974) and much of what I will present here is taken from that study.

Perhaps the most revealing single study was that of Kisslinger et. al. (1961). These authors made careful measurements of the SH radiation from small explosions in soil. In one experiment they found nearly identical SH radiation from a series of three shots in which the first 0.5 lb charge was tamped in soil at a shallow depth which led to cratering. The hole was then dug out into a smooth hemisphere and filled with water in which another 0.5 lb charge was detonated; the hole was filled again, and an 0.5 lb charge again detonated.

Notice the large number of causes of SH motion that these experiments eliminate: There can be no prestress or tectonic strain release in water; and anyway we would not expect the result to duplicate time after time. It could not be due to a truly random Taylor instability (Wright and Carpenter, 1962) because the pattern would be different from shot to shot. It could not be due to cracking and the consequent conversion because water and soil will not crack.

The only remaining possibility, it seems to me is that of anisotropic properties of the soil leading to greater work being performed by the source in one direction as compared to another. A force combination of this type may be characterized in moment tensor language as a compensated linear vector dipole and would have neither a double couple or an explosive source mechanism.

This hypothesis would explain why nearby shots in Yucca can have different "tectonic" generation; ie their soil properties are different due to different sedimentary layers at shot point.

Additional experimental work in the laboratory, in the field with small explosives, and at Yucca might be justified along these lines. Another study would be to analyze the event Piledriver, both surface and body waves to see if a vector dipole will match better than a double couple.

The classic seismological approach to the problem was that which culminated in the study by Toksoz and Kehrer (1972) in which they estimated the proportion of the "tectonic" release for a number of NTS and Amchitka explosions. They concluded that the tectonic release was consistent with a strike-slip mechanism and that the effect of the additional rediated energy would not change M_s more than 0.1 magnitude units.

Rivers (1980) has recently re-analyzed the Piledriver event (the NTS event with the largest proportion of Love to Rayleigh amplitude) using a moment tensor approach, and has concluded that the equivalent best fitting double couple is a thrust event which would lower the observed M_s of the explosion..

The question of whether the "tectonic" component will increase or decrease M_s is complicated. If the tectonic source is strain release or a triggered earthquake then the M_s may decrease or increase depending on details of the radiation pattern. If the source is due to conversion at cracks, then perhaps the energy will be subtracted from that originally coupled and the LR will decrease. On the other hand, perhaps cracking or block sliding increases the overall coupling. Clearly it is difficult to reach a conclu n.

Finally, if anisotropy is the cause a as suggested at the beginning of this report then a detailed theoretical st. is called for, but intuitively I picture an expanding ball of pressure whi. does a fixed amount of work. Then if some of this work goes into LQ, then . the average the M_s would be decreased.

Rygg (1979) has pointed out that for an event at Semipalatinsk the Rayleigh wave is reversed relative to other events, and that the signal is in addition delayed by 4-6 seconds and has a substantial Love wave component. Because of the delay Rygg concludes that the change in polarity is due to spall. However, spall does not explain the existance of large Love waves. Another possibility is that the reversed polarity shots are sufficiently shallow that their polarity is reversed as discussed by Gupta and Kisslinger (1964). Along with a shallow shot we may imagine increased cracking or greater anisotropy which will generate the Love waves. It does, however, remain to explain the 4-6 second delay. It may be that the "tectonic" source function is delayed by this amount of time for various physical reasons. It is also not completely clear that a full theoretical discussion of the problem would not show this delay to be normal. To my knowledge the question of a delay has never been investigated at NTS, although von Seggern (19/2) has used the cross-correlation between LR waveforms to locate NTS events. Investigation of the relative delays for selected pairs of NTS events would seem to be a fruitful area for study.

SELECTED REFERENCES

Blandford, R. and D. Clark, (1974). Detection of long-period S from earthquakes and explosions at LASA and LRSM stations with application to positive and negative discrimination of earthquakes and underground explosions, TR-74-15, Teledyne Geotech, Alexandria, Virginia.

Gupta, I. N., and C. Kisslinger, (1964). Model study of explosion-generated Rayleigh waves in a half-space, BSSA 54, 475-484.

Kisslinger, C., E. J. Mateker, Jr., and T. V. McEvilly, (1961). SH motion from explosion in soil, JGR, v. 66, 3487-3496.

Rygg, E., (1979). Anomalous surface waves from underground explosions, BSSA, 69, 1995-2002.

Toksoz, M. N. and H. H Kehrer, (1972). Tectonic strain release by underground nuclear explosions and its effect on seismic discrimination, Geophys, J. Roy. Astr. Soc., v. 31, 141-161.

von Seggern, (1972). Relative location of seismic events using surface waves, Geophys. J. R. A. S., 26, 499-513.

Wright, J. K., and E. W. Carpenter, (1962). The generations of horizontally polarized shear waves by underground explosions, J. Geophys. Res. v. 67, 1957-1963.

California Institute of Technology, Pasadena
J. Bernard Minster

Seismic Field Determination

Anomalous Radiation Due to Imperfect Bonding at Interfaces

key words: Anomalous radiation
S wave generation
Tectonic release
Block motion

This report draws attention to the possibility that block motion due to imperfectly bonded interfaces may contribute to the generation of anomalous radiation by buried explosions. The mechanism is reasonably efficient only if certain conditions are met, but there is reason to believe that this cannot be ruled out at the present time.

The classically used boundary condition at the interface between two elastic half-spaces involves continuity of tractions and displacements. The simplest generalization which does not depart from linearity consists in allowing shear slippage along the boundary. The shear traction is then assumed to be porportional to the velocity jump across the interface. The problem which has been studied so far involves a monopole (explosive) source near such an interface (Salvado and Minster, 1980). The most important conclusions are:

1) Imperfectly bonded interfaces permit the conversion of P waves to S waves even in the absence of an impedance contrast across the interface (Murty, 1975, 1976)

2) Interaction of the radiation field of a buried explosion with such an interface will generate anomalous shear radiation even in the absence of preexisting tectonic stresses (compare with the results of Kisslinger, et al., 1961)

3) The conversion of P waves to S waves may be compared with similar conversion at the free surface; it is most efficient when the interface is debonded and when the source is close to the interface.

4) It is expected that the true rheology at interfaces (joints, block boundaries) near a buried explosion is nonlinear. However, comparison with the few cases when realistic boundary conditions have been successfully studied (e.g. Miller, 1979) shows that linear viscous bonding may be a good approximation for the case of large amplitude, high frequency incident waves.

5) The SH radiation pattern associated with a <u>vertical</u> slipping interface near a buried explosive source is superfically similar to that of a double-couple. It differs considerably from a quadrupole in detail, particularly since the phase pattern (first motion) is not that of a double-couple. However, at present, it would seem that distinguishing between both cases is difficult with limited data.

Discussion:

We have only solved the body wave problem, in the high frequency, first motion approximation. The next logical steps in this investigation are considerably more difficult. The following approach may be suggested, not necessarily in the proposed order:

1) Identification of one, or several, instances when SH waves from a buried explosion possess a radiation pattern compatible with this mechanism.

2) Theoretical investigation of the case of a family of parallel steeply dipping interfaces.

3) Excitation of surface waves by this mechanism.

4) Effects of finiteness of the slipping joint, and associated propagation of cracking along joints.

5) Effective rheology of a material with a family of such joints, and assessment of their effect on near field waves (linear and nonlinear elasticity and anelasticity).

Steps are currently being taken to investigate the first two items of this list, and to attempt to define an equivalent transparent source of radiation.

Conclusion:

Based on our results to date, we think that block motion, as modelled through the linear mechanism described here, has a potential influence on the partitioning of energy between various seismic phases. In that respect, it may have some influence on seismic yield estimates.

It must be emphasized that in the absence of tectonic prestress, no additional energy is liberated by this mechanism and added to the radiation field. This is a purely passive mechanism which can only absorb energy, and modify the relative partitioning between P waves and S waves. It is not known at present whether significant energy can be removed from body waves and transferred into surface waves in this fashion.

J. B. Minster

Bibliography

Kisslinger, C., E. J. Mateker, T. V. McEvilly (1961). SH motion from explosions in soil, J. Geophys. Res., 66, 3487-3496.

Miller, R. K. (1979). An estimate of the properties of Love-type surface waves in a frictionally bonded layer, B.S.S.A., 69, 305-317.

Murty, G. S. (1975). A theoretical model for the attenuation and dispersion of Stoneley waves at the loosely bonded interface of elastic half-spaces, Phys. Earth Planet. Int., 11, 65-79.

Murty, G. S. (1976). Reflection, transmission and attenuation of elastic waves at a loosely-bonded interface of two half spaces, Geophys. J. R. astr. Soc., 44, 389-404.

Salvado, C., J. B. Minster (1980). Slipping interfaces: A possible source of S radiation from explosive sources, B.S.S.A. (in press).

University of Colorado, Boulder

11. TECTONIC GENERATION

by

Charles B. Archambeau

Introduction

Our current understanding of "tectonically generated" seismic effects is that whenever the medium is stressed in any manner (overburden, tectonic stresses, etc.) an explosion which creates a cavity and shatter zone will induce an "anomalous" motion in the medium, which is due to the relaxation of the initial stress in the medium around the fracture zone. Anomalous seismic radiation is defined to be that part of the seismic field that does not arise solely and directly from the isotropically occurring conversion of the explosive shock wave into an elastic compressional wave. That is, any part of the field not corresponding to a pure isotropic compressional source.

Not all of the observed anomalous radiation from explosions need arise from stress relaxation effects, since anisotropy and other local inhomogeneities (including the free surface of the earth) can give rise to similar effects. Kisslinger (1976) and Bache (1976b) have reviewed the likely processes for production of the observed anomalous radiation. It is doubtful that the seismic perturbations due to effects other than "tectonic" are nearly sufficient to account for the observed anomalous radiation (see Toksoz and Kehrer, [illegible]; Archambeau, 1972; Bache, 1976a); especially the anomalous radiation associated with long period surface waves.

Physical Processes of Stress Relaxation

The explosion induced stress relaxation has generally 41
been associated with <u>tectonic</u> stress relaxation, but it should

41. Charles J Archambeau
"Tectonic Generation"

42. Thomas C. Bache
"Tectonic Generation"

43. David G. Harkrider
"Tectonic Generation"

44. Thomas D. Eisenhauer
"Tectonic Effects"

Surface Wave Attenuation

45. Keith Priestley
"Effects of Attenuation on Surface Waves"

46. Otto W. Nuttli
"Effects of Attenuation on Surface Waves"

47. Brian J. Mitchell, St. Louis University
"Effects of Attenuation on Surface Waves"

48. Thomas C. Bache
"Effects of Attenuation on Surface Waves"

TABLE 1
List of Contributions

Body Wave Magnitude

1. James Taggart and E.R. Engdahl, U.S. Geological Survey
"Routine Determination of Earthquake Magnitudes by the USGS National Earthquake Information Service (NEIS)"

2. Robert J. Zavadil, Air Force Technical Applications Center (AFTAC)
"Body Wave Magnitude"

3. Karl F. Veith, Teledyne Geotech, Garland, Texas
"Body Wave Magnitude"

4. Thomas D. Eisenhauer, AFTAC
"Body Wave Magnitude"

5. Robert R. Blandford, Teledyne Geotech, Alexandria, Virginia
"Body Wave Magnitude"

6. Thomas C. Bache,* Systems, Science and Software, La Jolla, California
"Definitions of Body Wave Magnitude"

7. Otto W. Nuttli, St. Louis University
"Estimation of Body Wave Magnitude"

Source Coupling

8. Lane R. Johnson, University of California, Berkeley
"Near-Source Effects on P Waves"

9. Donald L. Springer, Lawrence Livermore Laboratory, Livermore, California
"Body Wave Coupling"

10. Robert R. Blandford
"Experimental Data on Body Wave Coupling"

11. John R. Murphy, Systems, Science and Software, Reston, Virginia
"Experimental Data on Body Wave Coupling"

12. William R. Perret, Albuquerque, New Mexico
"Evaluation of Body Wave Coupling Through Experimental Data"

13. Donald V. Helmberger, California Institute of Technology, Pasadena
"Near Source Effects on P-Waves"

14. John G. Trulio, Applied Theory, Inc., Los Angeles, California
"Seismic Yield Determination"

*Now at DARPA, Arlington, Virginia

be emphasized again that any stress in the solid medium, whatever its origin, will cause a radiation effect (e.g., lithostatic stress). It is generally agreed, however, that stresses of tectonic origin are most likely to be responsible for the larger effects observed. What is currently less clear is whether the stress relaxation is due to the creation of a nearly spherical shatter zone (e.g., Archambeau, 1972; Archambeau and Sammis, 1970), or whether it is largely due to "triggering" of an earthquake – that is associated with induced faulting corresponding to material failure having a strongly asymmetrical, or linear, pattern (e.g., Aki and Tsai, 1972). In the latter case, shatter zone induced relaxation occurs, but in addition failure along a preexisting, or newly created, long linear fracture is also thought to occur and would generate additional "anomalous" radiation, especially at the lower frequencies. The distinction between these mechanisms has importance for discrimination as well as yield estimation in that, if stress relaxation due to triggered faulting occurs, then much larger perturbation of the low frequency radiation would be expected than would be the case for shatter zone induced radiation. For triggering then, one might expect perturbations that would make accurate yield determinations using M_s highly uncertain – for example, the surface waves from triggering of a modest sized thrust earthquake could easily completely cancel or overwhelm the explosion generated surfaces at all azimuths in the period range around 20 seconds. Further, the M_s value for the explosion plus earthquake could, in many cases, be earthquake-like, making discrimination by m_b-M_s problematic. The size and nature of the effects of triggering would be difficult to accurately predict or to correct for, since not only would information regarding stress drops be necessary, but knowledge of the location and orientation of the fault plane and rupture rate would be required. On the other hand, if the physical mechanism is

simply stress relaxation around the roughly spherical shatter zone created by the explosion, then it is likely that the perturbations in the seismic radiation would be relatively smaller at low frequencies and more easily predicted and corrected for; in that the origin and geometry of the failure zone, as well as its formation rate, would be known quite well. Further, the size of the shatter zone is predictable and related to the explosion yield. (Prediction of the size, or radius, of the shatter zone is not normally available from either scaling law or empirical data, or from numerical code calculations. However, it could be obtained in these ways. In this regard, see the recommended research section below.) Finally, the tectonic and/or lithostatic stress levels in the shallow depth range for average explosions can be reasonably estimated, especially if earthquakes from the area can be studied, and a correction for stress relaxation effects on surface wave observations could then be made, probably with reasonable confidence. (Such a correction would, of course, entail uncertainties and surface wave based yield estimates would have to be compared with body wave yield estimates.) In any case, research would be required in order to develop the basis for any such correction and to determine its accuracy.

Of the two possible mechanisms for tectonic release, it seems most likely that both processes have occurred in the past. However, the simple spherical shatter zone relaxation process has probably been the mechanism giving the anomalous seismic radiation observed from most underground explosions, with triggering occurring for only a few events (about five percent or less of the total). This estimate is based on a variety of physical evidence (i.e., aftershock locations versus time, location of the "anomalous event", near field strain observations, etc.), plus the fact that the anomalous surface wave radiation can be adequately explained by spherical shatter zone induced stress relaxation for explosions with

F factors* less than around one. (When $F \approx 1$, the seismic energy of tectonic origin is about the same as that from the pure explosion.) Even for explosions where F factors considerably larger than one are observed, it is not certain that large scale faulting is required. In any case, it is likely that triggering is quite rare.

Characteristics of Observed Anomalous Body and Surface Wave Radiation from Explosions and Predictions from Explosion Induced Tectonic Release

Some typical examples of the nature of the anomalous radiation from explosions are shown in the following figures. The effects are most pronounced for surface waves and in this discussion such effects are emphasized. Nevertheless, body waves are perturbed as well, but the effect on body wave magnitudes m_b are much less than the effect on the surface wave magnitude M_s; so long as m_b is measured from the first cycle of the P wave motion or is measured spectrally by narrow band filtering using the group arrivals within 1 to 1.5 seconds of the direct P wave first motion.

An example of the effects of tectonic release on the first arriving compressional waves from an explosion is shown in Figure 1 (Archambeau, et al., 1974). The event modeled is Handley (1.1 megaton) at a teleseismic distance. The tectonic release mechanism modeled is stress relaxation around the roughly spherical explosion generated shatter zone, where the initial prestress is taken to be 65 bars and homogeneous. This is a modest prestress level, and levels of from two to three times this value would not be unlikely. The orientation of the prestress was pure shear in the

---

*An approximate relation between the F factor, used to characterize the anomalous, or tectonic, component of the radiation relative to the direct explosion component is (Toksoz, et al., 1965): $E_{tect.}/E_{expl.} = 4/3\ F^2$.

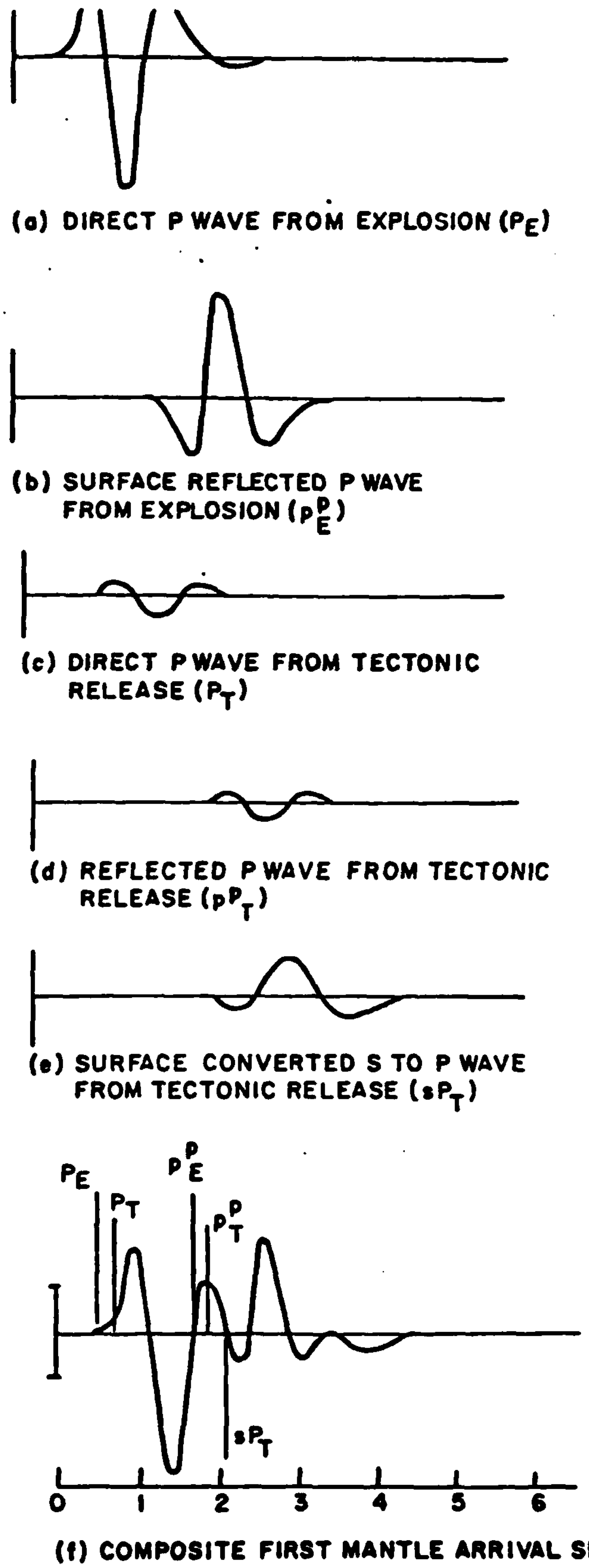


Figure 1. Theoretical compressional (P) wave phases generated by the Handley explosion and associated tectonic release. The Handley event (1100 kt) modeled by the explosion source model T-1. Prestress (σ_{13}) for tectonic release taken to be 65 bars, shatter zone radius R_0 = 750 meters. Earth structure CIT 109--Low Q model. Distance 4066 km, azimuth 30°. Vertical component LRSM short period seismometer, source depth 1.2 km.

horizontal plane, so that relaxation of stress around the shatter zone is equivalent to a strike-slip double couple point source. For a homogeneous prestress, the stress relaxation around a spherical shatter zone is such as to always produce pure quadrupole radiation - that is a simple double couple equivalent. The figure shows the time domain pulse contributions to the overall P wave train and the important point is that the direct P wave from tectonic release is much smaller than the explosion generated P wave. On the other hand the tectonic S wave is comparable to the explosion P wave, but only contributes upon reflection at the free surface so that it influences the wave train at later times. The ratio of P to S wave production by tectonic release of this type scales is:

$$\frac{A_p}{A_s} \propto \left(\frac{V_s}{V_p}\right)^3$$

where A_p and A_s are the amplitudes of tectonic P and S waves, and V_p and V_s are the P and S wave velocities in the source region. In general then, we expect the tectonic P wave to be of the order $3\sqrt{3}$ down in amplitude from the S wave. If we consider explosions with F factors near unity, so that the tectonic energy released is nearly the same as the explosive energy converted to seismic radiation, such as in the example in Figure 1, then the tectonic P wave will be five times smaller than the direct explosive P wave (i.e., the energy of both sources is about the same but 80% of the tectonic source energy is contained in the S waves produced). The example in Figure 1 illustrates this relationship. Thus, as noted earlier, if the body wave magnitude measurement is confined to the first cycle of P wave motion, then the effects of tectonic release will be minimized. A similar conclusion was reached by Bache (1976a) from a series of modeling experiments in which P waves from a number of explosions were studied, with tectonic release effects included to achieve detailed fits to the observations.

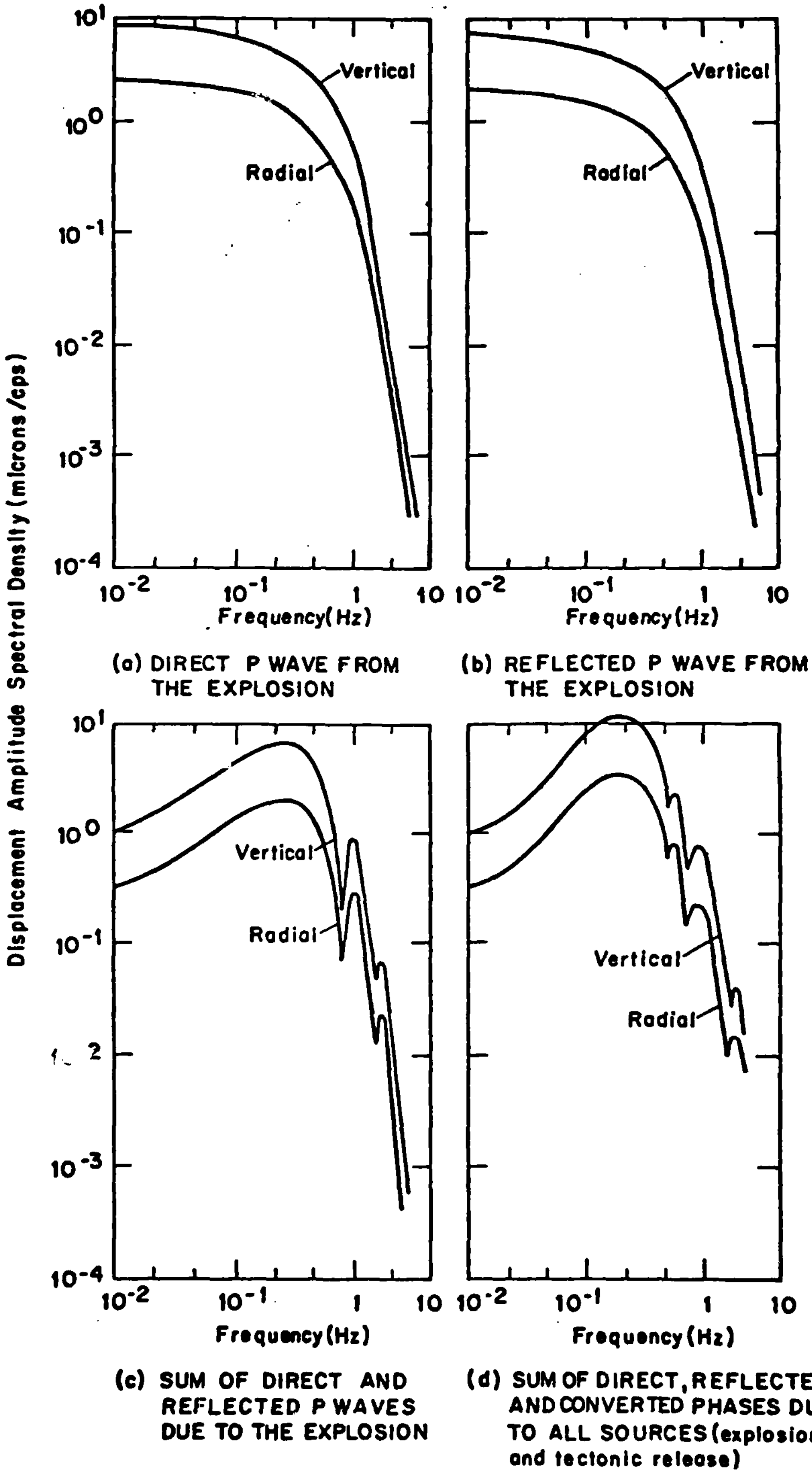


Figure 2: Theoretical spectra of explosion generated compressional arrivals from the Handley event at a distance of 4066 km and azimuth of 30°. CIT 109 Low Q model, explosion model T-1, source depth 1.2 km. Vertical and radial refer to the two components of the vector field at the earth's surface with radial denoting the horizontal component in the radial direction from the source point.

Figure 2 shows the spectra of the waves represented in Figure 1, with the last inset (d) showing the spectral perturbation resulting from tectonic effects. It should be emphasized that the composite spectrum shown in (d) is for the entire P wave train. If spectral methods are used, say narrow band filtering methods, to obtain a spectral magnitude for the first arrival pulse (i.e., the first cycle of the P wave train), then the result obtained would look like inset (c), rather than (d). Figure 3 shows the character of the entire predicted P wave train, including multiple mantle arrivals to be expected in the distance range near 4000 km, for the explosion above compared with explosion plus tectonic release. Finally Figure 4 shows a comparison with an observation, from the underground test Bilby. The predicted seismogram for the P wave train is remarkably similar to that observed, and this kind of agreement is not unusual. It is clear however, that the effect of tectonic release on the P waves is not large and that an explosion by itself could fit the observations adequately, especially when uncertainties in structure and the predicted explosive source function itself are taken into account. This is also illustrated in Figure 3 by the small differences between the theoretical seismograms with and without tectonic release.

The predictions of seismic radiation from tectonic release have usually assumed a uniform prestress condition in the medium prior to the creation of the explosive shatter zone, with the exception of the treatment employing a stress relaxation cut-off at some radius (R_s) in order to approximate the effects of a stress concentration in the medium (e.g., Archambeau, 1970, 1972). When a uniform prestress (extending to infinity) is used and a spherical shatter zone is created, then the radiated field is pure quadrupole and its far field spectrum is flat from zero frequency to a corner frequency,

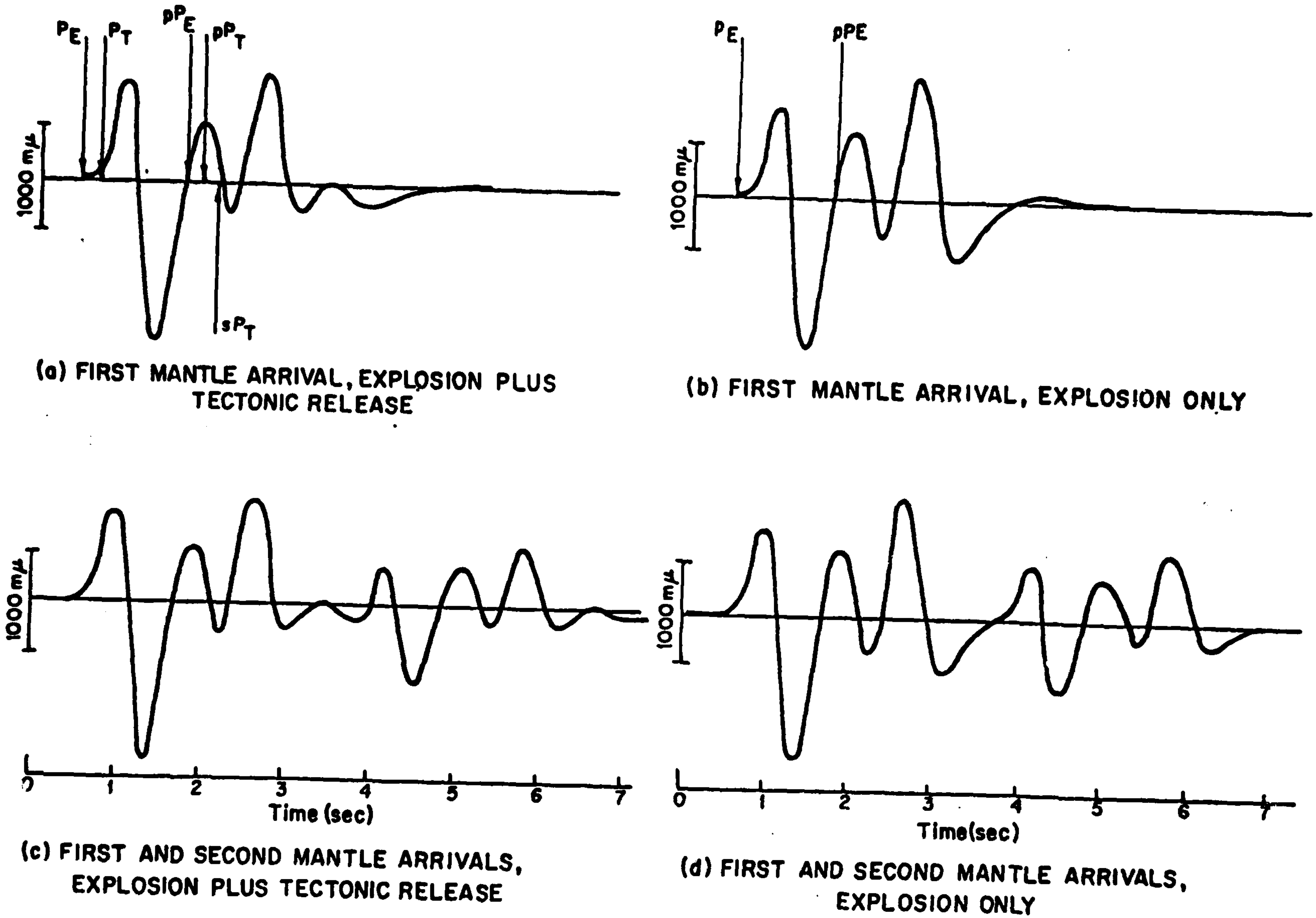


Figure 3: Comparison of theoretical signals at 4066 km from Handley (1100 kT) explosion model with and without tectonic release. Explosion model was T-1, tectonic release based on 65 bar pure shear prestress (σ_{13}) and shock generated shatter zone of 750 meters in radius, source depth 1.2 km.

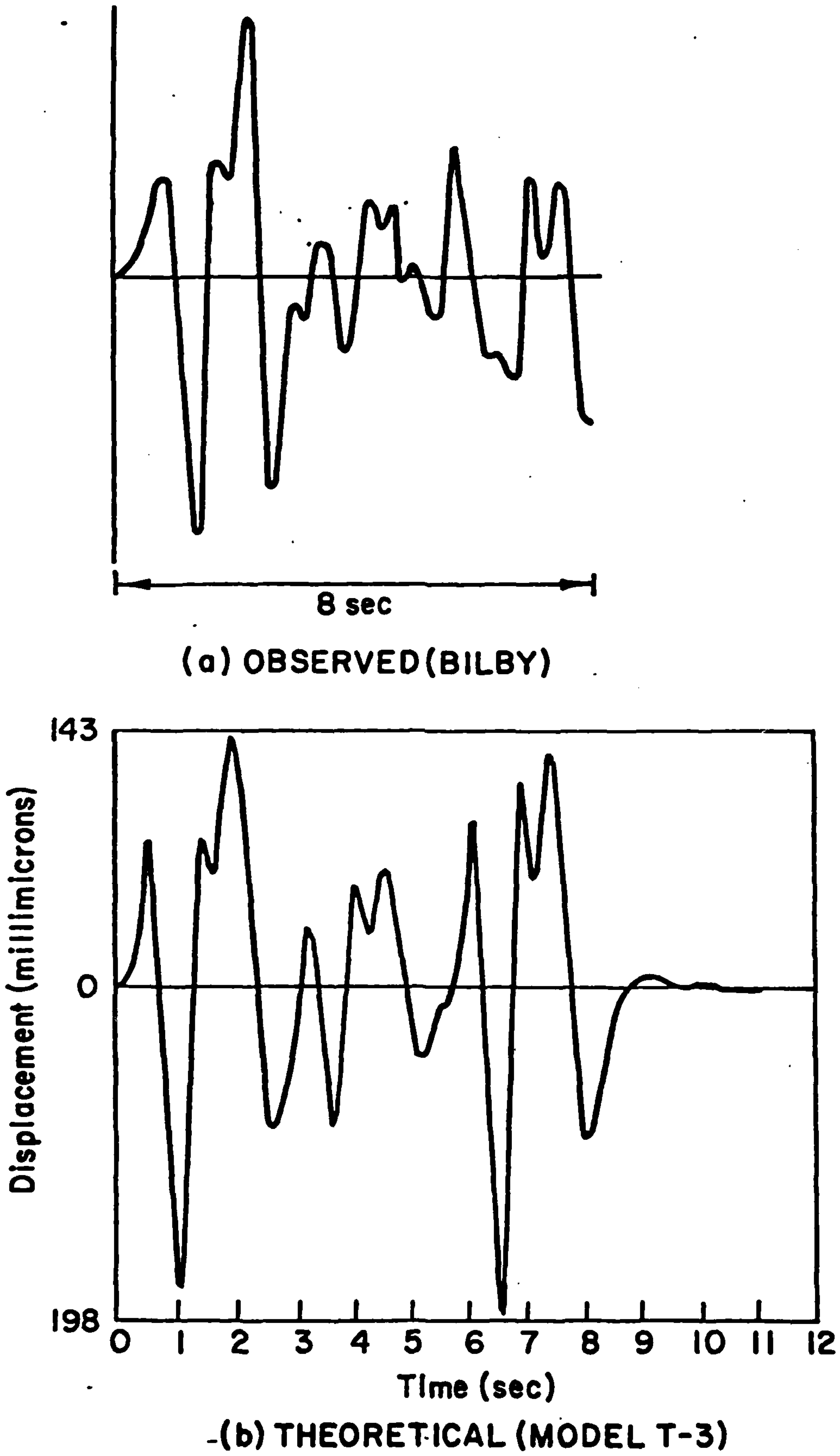


Figure 4: Comparison of observed and theoretical seismograms for the Bilby event at HNME, 4066 from ground zero.

beyond which it falls off as $1/f^2$ or $1/f^3$ with increasing frequency, depending on the rate of formation of the shatter zone. The previous examples, in fact, use the simplification of homogeneous prestress in an infinite medium (R_s infinite). When the "R_s approximation" is used however, the tectonic relaxation is confined to a finite volume and hence is less effective as a long wave length radiator. In this case the field is again quadrupole but the spectrum is peaked and is different from the uniform case in that the far field spectrum decreases to low frequency, rather than remaining flat. The implication here then is that for inhomogeneous prestress situations, or when a finite region of prestress is considered (obviously the earth is finite so that the prestress cannot be uniform to infinity and so a finite zone of stress relaxation is clearly required in any case), then the spectrum will be strongly peaked at rather high frequencies and the contributions to high frequency body waves and higher mode surface waves may be larger than expected. Thus in reality we might expect stronger perturbations in the body wave radiation, when strong stress concentrations are present, than were previously suggested. This could have an effect on the reliability of m_b estimates for both yield estimation and (even) discrimination.

The "R_s approximation" was a subject of considerable argument in the past, since, among other things, it contradicted results from other (kinematical) source models. (e.g., It implied that earthquake spectra should be peaked as well.)

Recently Stevens (1980) has considered the stress relaxation problem for a spherical shatter zone in an inhomogeneously prestress medium and succeeded in obtaining an exact solution for any arbitrary prestress condition, with the solution also accounting for all scattering from the spherical inclusion. The solution shows that the tectonic release

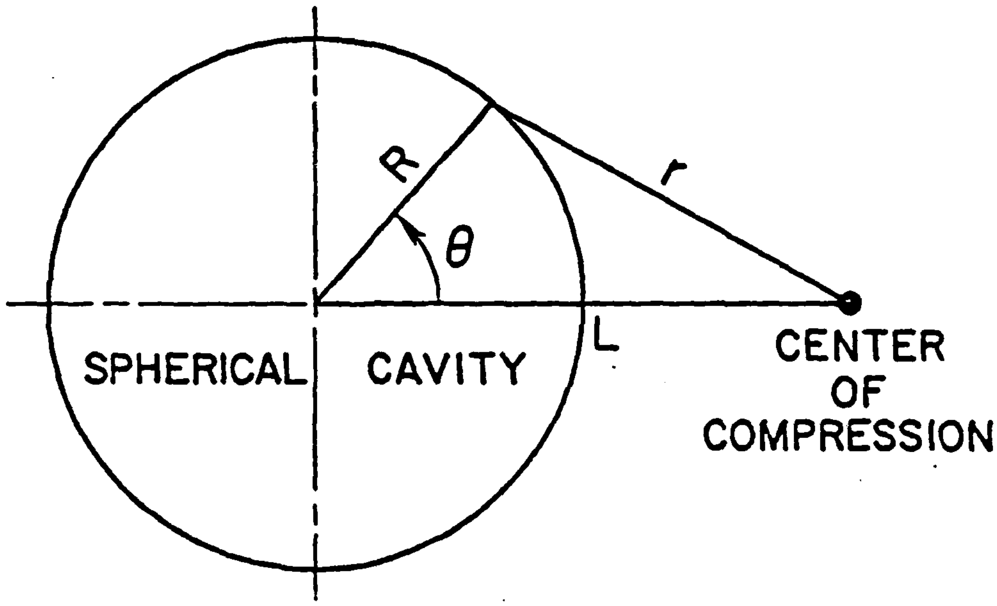


Fig. 5. Coordinate system used when a cavity is created near a preexisting center of compression.

radiation spectrum can, in fact, be strongly peaked when the prestress is concentrated near the explosion produced shatter zone. However, the solution, as expected, is more complex than is indicated by the "R_s approximation" for prestress concentration. In particular, it shows azimuthal dependence of the spectral peaking, with the spectrum nearly flat in some azimuths and very strongly peaked in others and the variation between these forms being dependent on the location and type of stress concentration. In addition, shifts in the corner or peak frequency toward higher frequencies are also predicted and are such that the corner or peak frequency is more a measure of the dimensions of the stress concentration than of the shatter zone dimension. Further, the radiation patterns and first motions for high frequencies near and above the peak, or corner frequency, are usually not quadrupole in form.

These results are illustrated by the following figures showing the effects associated with two kinds of initial stress concentrations. In particular, stress relaxation due to creation of a cavity in the vicinity of a center of compression within the medium and, as a second example, in the vicinity of a static point dislocation, clearly illustrate the spectral and radiation pattern effects described. The spatial relationship between the cavity and the center of the compressional stress concentration is shown in Figure 5. Note that L denotes the distance between the center of compression (or position of a point dislocation in the second example) from the center of the cavity.

The seismic radiation produced by the creation of the cavity (or shatter zone) in the vicinity of a center of compression is largely quadrupole at low frequencies but is more strongly perturbed by higher order multipoles at higher frequencies. Figure 6 shows P and S wave spectra, and wave forms, of the radiation near quadrupole node lines in the

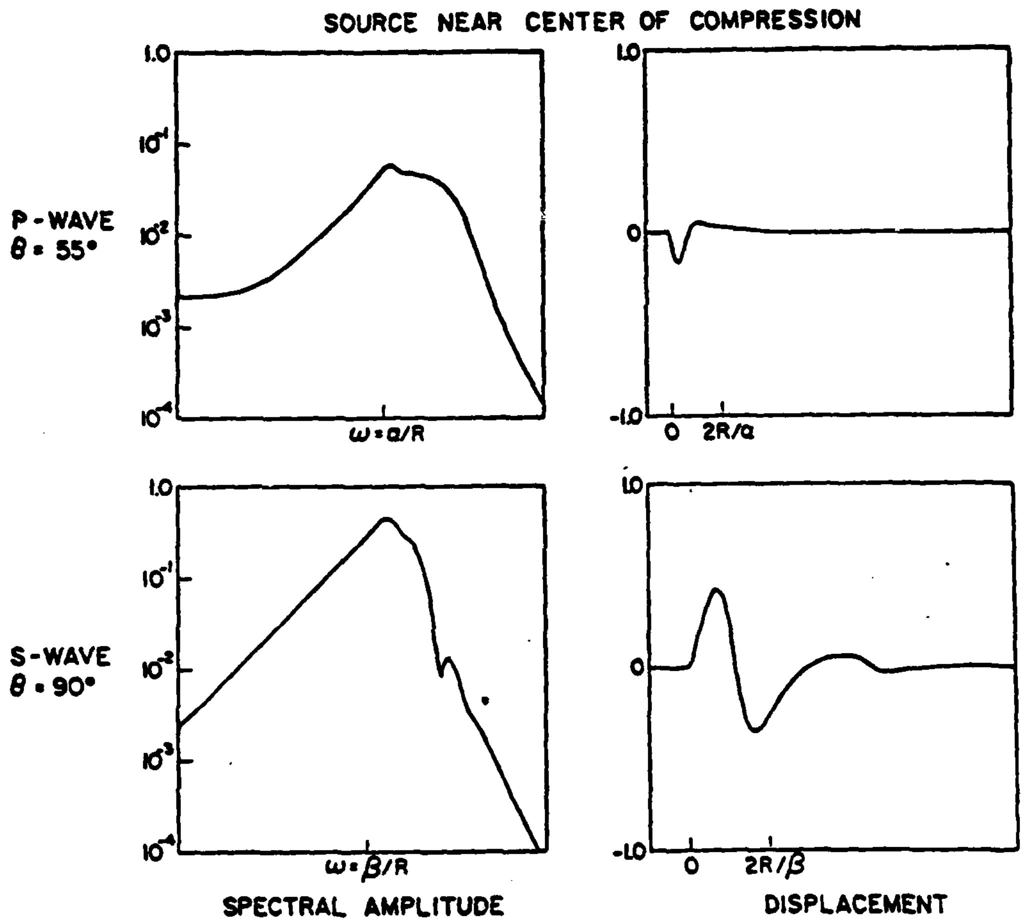


Fig. 6. Far field radiation near the quadrupole nodes. When the center of compression is located one radius away from the cavity, the radiation field is much like a pure shear field. Near the quadrupole nodes however, there is a substantial difference. The displacement does not vanish. It is reduced from the maximum by about a factor of 3. The pulse is oscillatory and the spectrum is peaked.

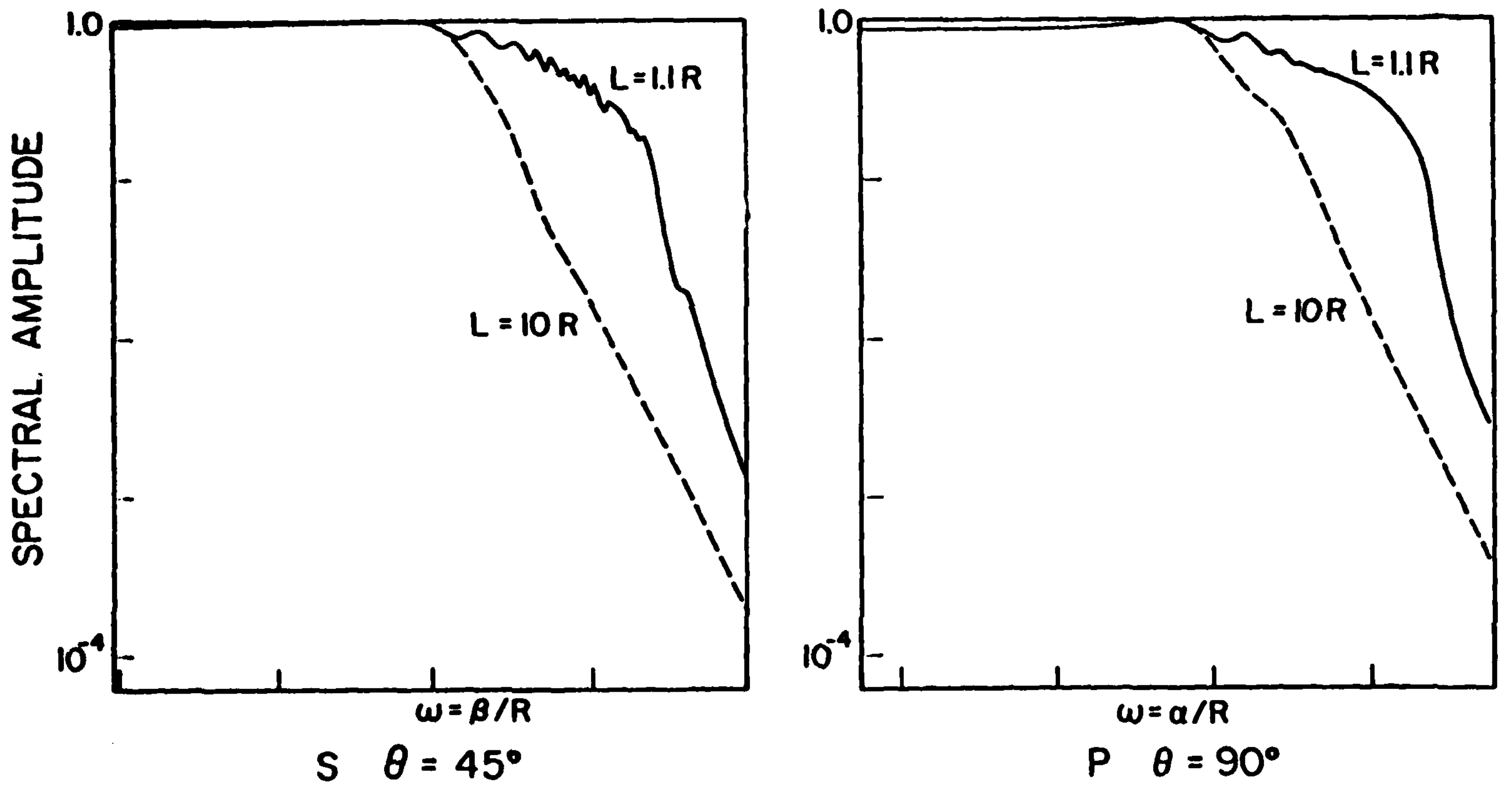


Fig. 7. Far field spectra near quadrupole maxima. The spectrum is strongly affected by the location of the stress concentration. Note the increase in corner frequency and the change in slope near the corner.

radiation pattern. Not only is the seismic radiation non-zero at the quadrupole nodes, the spectra is strongly peaked and the wave forms are correspondingly oscillatory. Further, the peaks or corner frequencies in the spectra are at higher frequencies when the cavity is close to the stress concentration than when it is further away. This effect is shown by the spectral plots in Figure 7. In this figure the spectra are normal and quadrupolar when the distance from the cavity to the compressional stress concentration is 10 cavity radii away ($L = 10R$). When the stress concentration is on the edge of the cavity ($L = 1.1R$), then the spectrum is considerably shifted towards higher frequencies. Further, the radiation is non-quadrupolar in the high frequency range. However, at the maximum in the P and S radiation patterns ($\theta = 90°$ and $\theta = 45°$ respectively) the spectra are only very slightly peaked, in contrast to the strongly peaked spectra (Figure 6) near the nodes in the radiation patterns.

The spectra vary considerably as a function of the type of stress concentration. Figure 8 illustrates the seismic radiation due to relaxation of stress when the initial stress is produced by a point dislocation near the shatter zone. Here the spectra are peaked and the wave forms oscillatory at all azimuths. The corner or peak frequencies are also at higher values than for pure quadrupole radiation from a homogeneous prestress.

Figure 9 shows the effects on P wave first motion, as a function of azimuth, for the two different types of stress concentrations. When the stress concentration, of either type, is far from the shatter zone ($L >> R$), then the pattern is quadrupole. For nearby stress concentrations the pattern is strongly perturbed and indeed is nonquadrupolar with higher order multipoles dominating the pattern.

Clearly, if there are strong stress concentrations in the vicinity of the explosion site then rather unpredictable

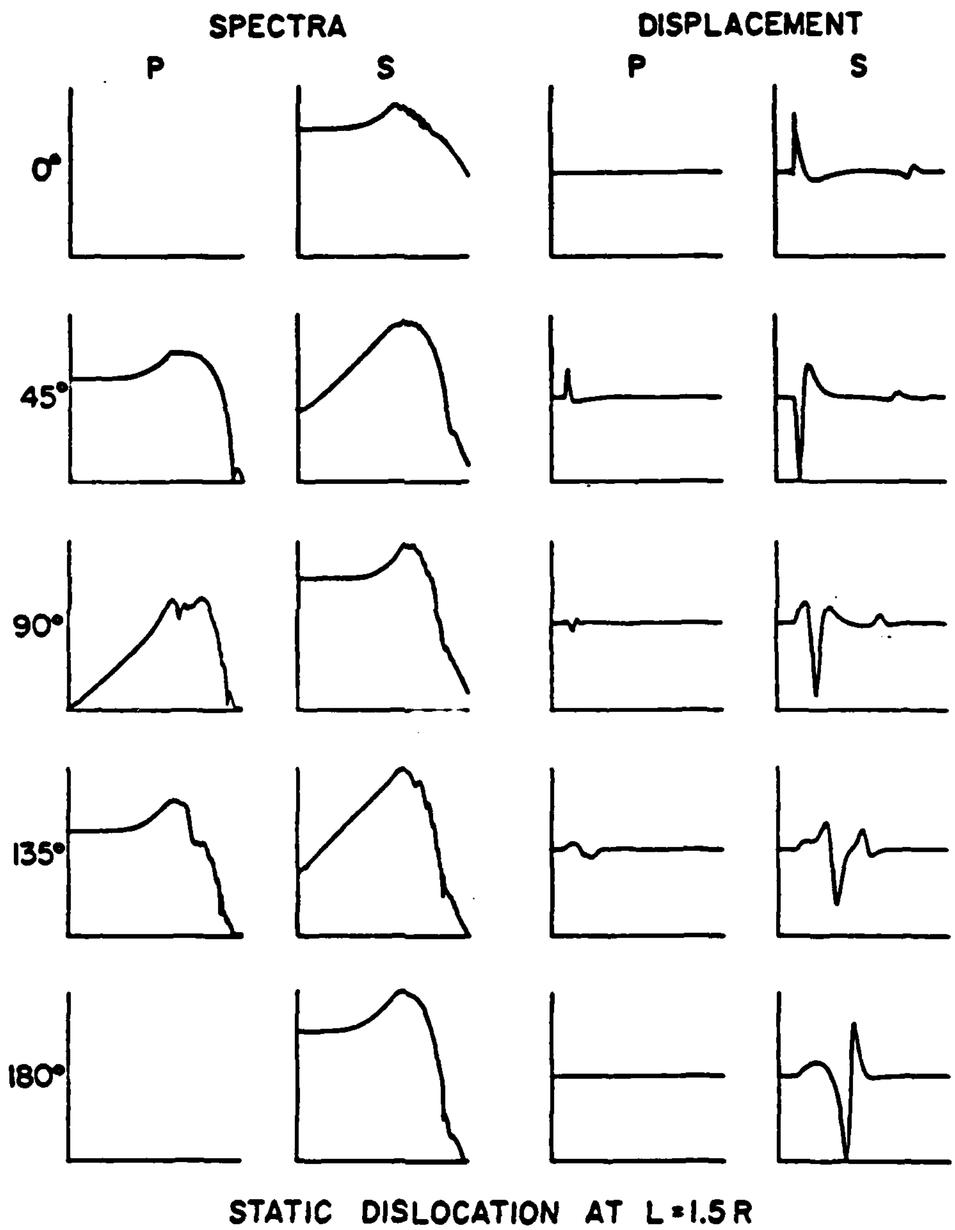


Fig. 8. Radiation field from spherical cavity created one-half radius from a point dislocation. Shown here are the far field spectra and waveforms as a function of angle around the cavity. The spectra are all strongly peaked. The waveforms are narrow on the side of the sphere near the stress concentration, broad on the other side. The second arrival seen on the first three shear waves is a diffracted phase which has travelled around the cavity.

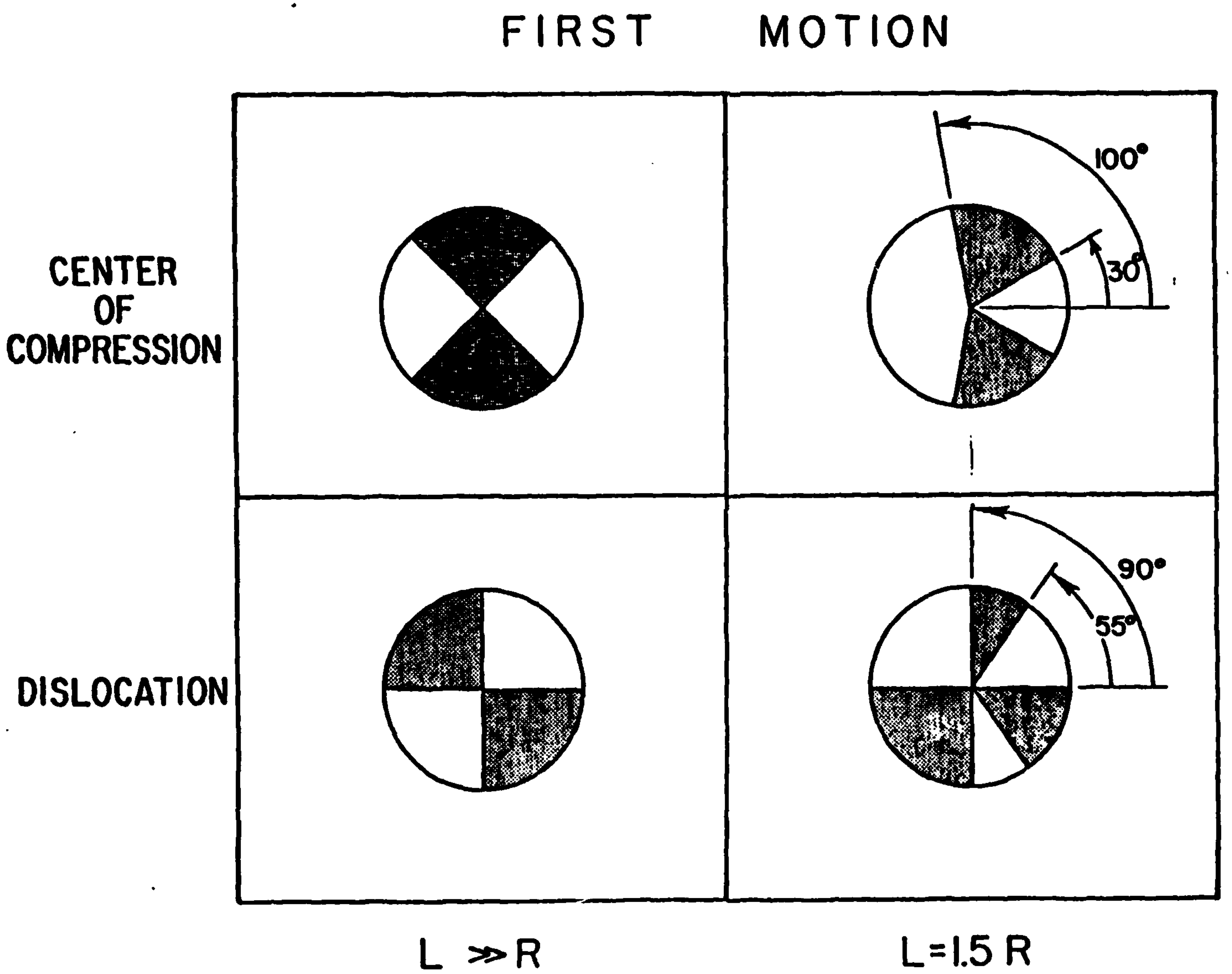


Fig. 9. First motions for localized prestress fields. For $L \gtrsim 2R$, the first motions are very nearly quadrupole in nature. At closer distances, the patterns shift. For the center of compression, there is simply a change in the "nodal" angles. For the dislocation a pulse directed opposite to the quadrupole pulse gradually becomes prominent at large angles. The angular distribution of first motions is less complex than the angular variation of spectra and waveforms.

perturbations in the high frequency P wave (and S wave) radiation can occur; since where and how large such stress concentrations might be,are hard to predict. However, there is really little doubt that stress concentrations of the size and character required to give "anomalous" stress relaxation effects can easily occur. For example, the mining operation necessary to conduct an underground test is sufficient to produce stress concentrations of the required size and type. (Consider the stress near an excavated cavity in an otherwise uniformly stressed medium, such as would be produced by simple gravity. Here the stresses near the excavation can reach kilobar levels if the test site is deep. If tectonic stresses and pre-existing medium inhomogeneities such as weak fault zones are also present, then the stress concentrations can be very large and have a complex spatial pattern.)

The way to minimize stress relaxation effects for compression wave measurements, whether "anomalous" or not, is, as previously mentioned, to make all measurements of magnitude, spectra, etc. on the first cycle of the P wave motion. When large stress concentrations are present, then the perturbations will be larger and more complex than if the prestress is uniform, but nevertheless will always be less than they are during later times in the P wave train. Estimation of how large the first and later cycle perturbations might be, for realistic cases, will require additional systematic study. However, such studies can clearly be carried out.

The effects of tectonic release on surface waves, as noted earlier, are relatively large and rather unambiguous. That is, large low frequency Love (SH) type surface waves are commonly produced by explosions, especially when detonated in tectonically active regions, and it appears that the only mechanism capable of explaining them is some form

of tectonic release (e.g., Archambeau and Sammis, 1970; Archambeau, 1972; Aki and Tsai, 1972; Toksoz and Kehrer, 1972). The only question seems to be precisely what mechanism of tectonic release is responsible, as noted earlier.

One line of evidence concerning the process involved in tectonic release is furnished by direct comparison of seismic radiation from an explosion and an earthquake occurring in the same region. One such pair, the Fallon earthquake (m_b = 4.4, depth 15 km) and the Shoal underground nuclear explosion test (m_b = 4.9; test medium, granite, F = .58) was studied in some detail by Lambert et al., 1972. These events were separated by only a few tens of kilometers so that the average tectonic environment, as measured by the long period surface waves, should be comparable. Figure 10 shows one of the striking differences between true earthquake long period radiation and the anomalous long period radiation from explosions. In particular, the ratio of Love to Rayleigh wave (L/R) spectral amplitudes strongly increases with increasing period for the earthquake, while the same ratio is essentially constant for explosions. The L/R spectral ration for the underground test Bilby (m_b = 5.8, test medium, tuff, F = .5) is also shown, for comparison with another (much larger) explosion event, and it also shows a very different period variation for L/R. The conclusion to be drawn is that the source of the anomalous explosion radiation, which is totally responsible for the Love wave excitation, must be quite different from a small earthquake, like the Fallon event. In particular it must be less efficient as a long wave length radiator than is a small earthquake and thus a smaller source dimension is implied. The tectonic stress relaxation induced by shatter zone creation is a source of relatively small characteristic dimension that produces quadrupole long period radiation with considerable SH wave production. Hence, it is a most likely model for the

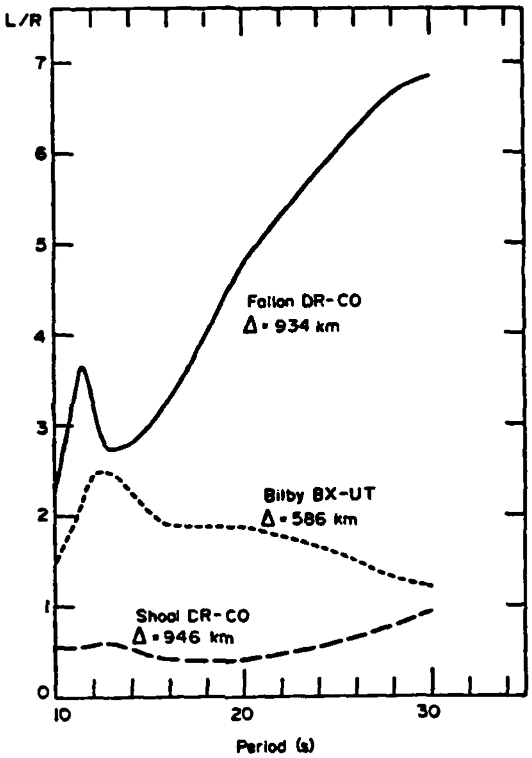


FIG. 10. The variation of L/R with period for the BILBY and SHOAL explosions and the Fallon earthquake, in the distance range near 800 km. L/R for the earthquake is larger than 2·0 and increases with increasing period. This difference in behaviour is a consequence of the much larger source dimension associated with the earthquake.

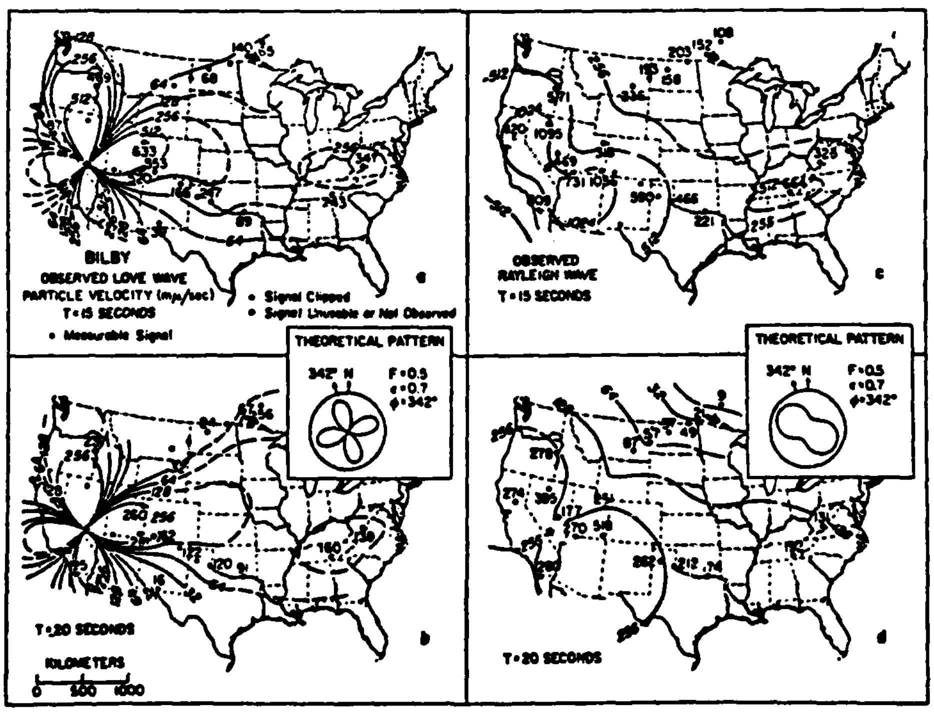


FIG. 11 Radiation patterns of Love and Rayleigh waves from BILBY at periods of 15 and 20 s. The insets show the theoretical pattern shapes obtained as a fit to the observations, using superposed monopole and quadrupole point sources with fixed relative excitation of quadrupole to monopole of $F = 0{\cdot}5$, with quadrupole 'strike' $\phi = 342°$. The point source equivalents are the same for both Love and Rayleigh waves. The factor $\varepsilon = 0{\cdot}7$ is the particle orbit ellipticity factor for the Rayleigh waves, which depends on the structure used in the theoretical calculations.

anomalous explosion source component. Figure 11 shows the form of the observed radiation patterns for Love and Rayleigh waves at periods T = 15 sec and 20 sec. The insets are the theoretically predicted surface wave patterns based on tectonic release due to the explosion produced spherical shatter zone, plus the pure explosion monopole field (Lambert et al., 1972). The prestress orientation used is such as to be consistent with the predominently strike slip mechanism associated with the Fallon earthquake, while the magnitude of the initial stress required to fit the amplitude of both the Rayleigh and Love waves was 65 bars (Archambeau and Sammis, 1970). The (mean) prestress level seems entirely reasonable. Thus the "shatter zone model" seems consistent with the observations of surface wave radiation, in that the values of prestress magnitude and orientation required to fit both Love and Rayleigh wave observations are compatable with the tectonics of the region.

Figures 12 and 13 show how well this model simultaneously fits the long period surface wave radiation for the Shoal and Bilby explosions. Here observed L/R ratios at several stations (where the maximum observed surface wave amplitudes are near 15 seconds) are compared to the predicted ratio at different azimuths. The agreement is very good, considering the probable lateral refraction effections to be expected. Similar analysis for numerous other explosions, for example by Toksoz and Kehrer, 1972, shows similar results.

Hence it would appear that the production of seismic radiation by spherical shatter zone induced (tectonic) stress relaxation can explain the anomalous long period surface wave observations, and the P and S wave train complications as well, in the great majority of observed cases. Particularly for the long period surface waves, which are relatively unaffected by small dimensional high stress concentrations, it should be possible to estimate the magnitude of the

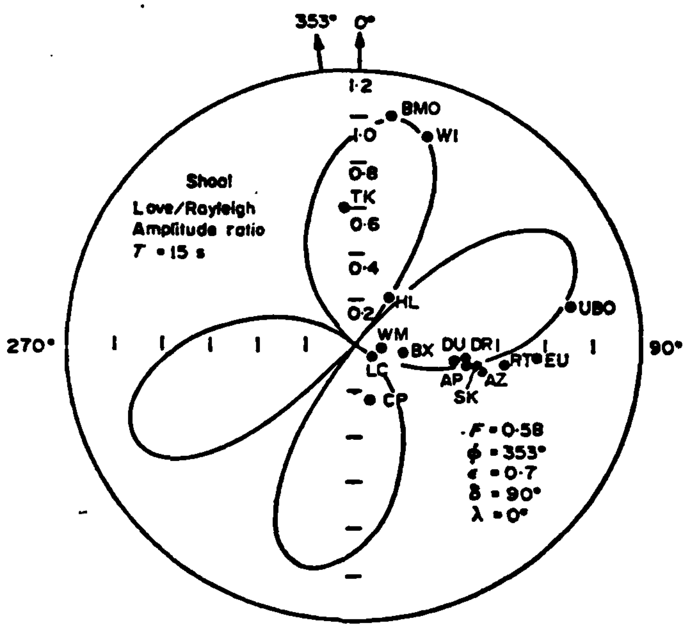


FIG. 12. The theoretical L/R ratio for SHOAL, at 15-s period, as a function of azimuth and using the source parameters indicated. Observations at various azimuths where signal-to-noise ratios where high are indicated and identified by abbreviated station symbols. This also shows the nature of the L/R azimuth variation (which is nearly identical to that for BILBY) and that a single equivalent source fits both Love and Rayleigh wave radiation for this event.

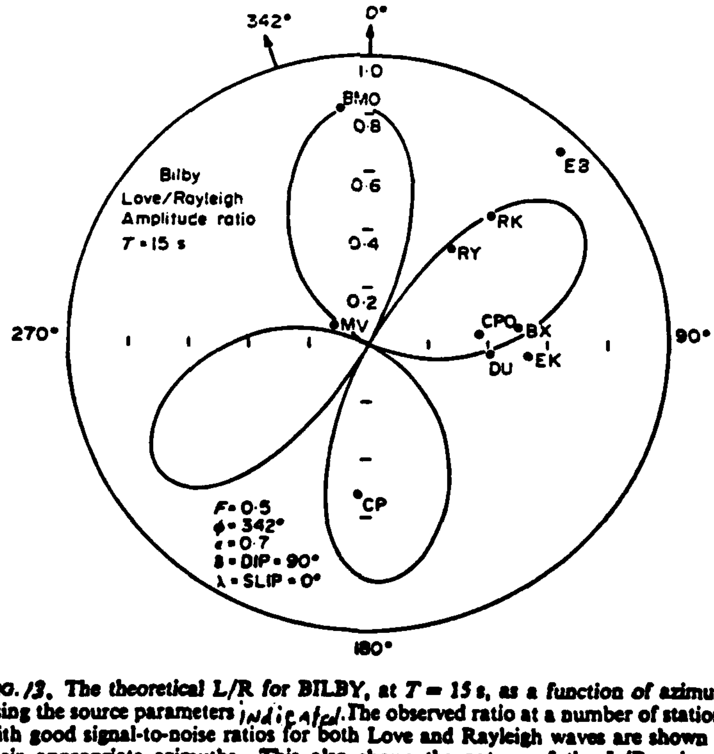


FIG. 13. The theoretical L/R for BILBY, at $T = 15$ s, as a function of azimuth using the source parameters indicated. The observed ratio at a number of stations with good signal-to-noise ratios for both Love and Rayleigh waves are shown at their appropriate azimuths. This also shows the nature of the L/R azimuth variation, and it is clear that a single equivalent source fits both Love and Rayleigh wave radiation for this event.

anomalous effects (perturbations) to be expected. On the other hand the body wave perturbations are much more difficult to predict, because of their dependence on details of the spatial dependence of the prestress, but the large effects can be avoided if only the first cycle of the P wave train is used in the event analysis.[*]

Figure 14 (Archambeau et al., 1974) shows, among other things, the effects of tectonic stress on m_b and M_s for explosions in granite, covering the yield range from 1 KT to 1000 KT, where shatter zone stress relaxation is assumed. The propagation path is appropriate to NTS, so that the upper mantle includes a well developed low velocity and low Q zone. For this reason the theoretical m_b values are somewhat lower than would be expected in a non-tectonic region, but are, on the average, what appear to be appropriate for NTS and the Basin and Range Geologic Province. The circles represent predictions of m_b and M_s values[**] from the explosions alone, without tectonic release. The measurements at these points are made from the synthetic seismograms, using the first cycle of the P wave train (denoted $m_b^{(1)}$) and the Rayleigh wave vertical component at 20 seconds. The "error bars" (upward vertical lines and horizontal lines attached to each circle) indicate the possible increase in M_s if the Airy phase is measured and, for m_b, the increase to be expected

---

[*] The situation for the prediction of anomalous effects in this context is similar to the situation described by Bismark concerning Prussia and Austria; to paraphrase: "The situation (for long period surface waves) is serious but not hopeless, the situation (for short period body waves) is hopeless but not serious."

[**] The superscript "R" is used on M_s in the figure to identify the M_s measurement as being associated with the Rayleigh wave. This was employed because a superscript "L" was used for the similar magnitude measurement from Love waves. This notation is unnecessary here as M_s will always be the standard magnitude from the Rayleigh wave.

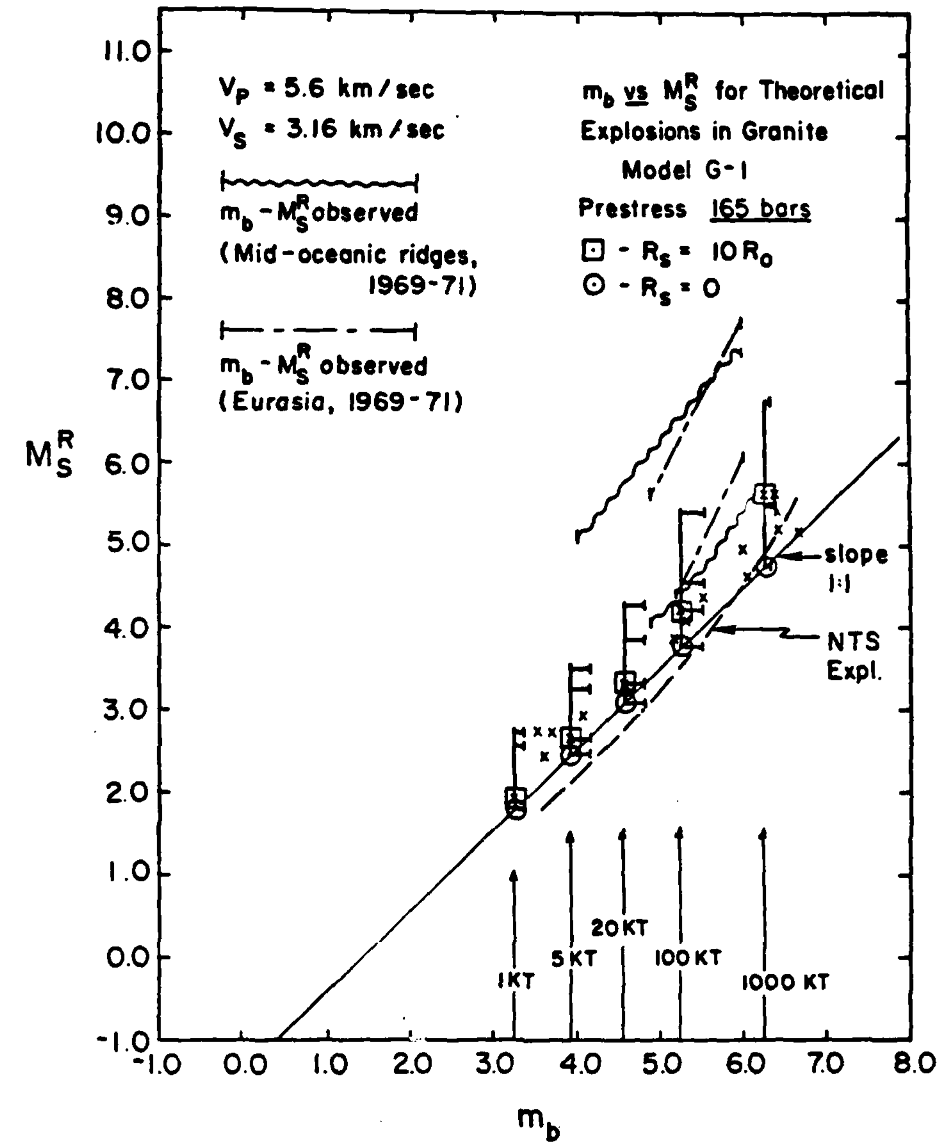


Fig. 14. Rayleigh surface wave magnitude as a function of body wave magnitude for explosions in granite (model G-1). The circles denote theoretical values of $m_b^{(1)}$ and M_s for zero prestress, the squares for 165 bar prestress. The horizontal lines indicate the range of possible m_b values, the vertical lines indicate the range of possible M_s values. The large variation in M_s arises from the possibility of measuring an Airy phase. The variation in m_b arises from the possibility of measuring the 2nd or 3rd cycle in the P wave train rather than the first. The upper and lower bound lines for earthquake data are shown for comparison along with the mean NTS explosion magnitude line. The X symbols are individual NTS explosions defining the extreme upper range of observed M_s values for explosions from Nevada. The yields of the theoretical explosions are indicated along the m_b scale.

if the larger of the second or third cycle in the P wave train is used to measure m_b. The squares correspond to the m_b and M_s values that would be obtained with 165 bars prestress. The M_s value here is not, however, an average over the whole radiation pattern, but the largest M_s that could be measured. It is therefore an upper bound on the (positive) perturbation in M_s that can occur for this kind of tectonic stress release. Naturally the "error bars" are dragged along with the point in the m_b - M_s plane and the upper horizontal bar is associated with the limit M_s measured from the Airy phase. The horizontal extent of the bar indicates the perturbation possible in m_b due to "cycle selection". The prestress orientation was taken to be such that the radiation quadrupole corresponded to a "strike-slip" equivalent double couple. For other choices of prestress orientation, the maximum, and average M_s as well, could be much reduced instead of being increased. (e.g., A thrust equivalent could probably cancel or even reverse the directly generated explosion Rayleigh waves.) In any case the figure illustrates the size of the effects for a particular case, and they are significant for M_s. They are not very large for m_b and indeed the variations in m_b don't change much with prestress in this special case. However, the calculation was done with essentially uniform prestress and when stress concentrations are present, as they almost certainly would be in actuality, the m_b variations could be much larger. Note also that the $m_b^{(1)}$ values are essentially identical with and without prestress, as shown in the figure. (i.e., The circle and square points are only displaced vertically, showing only a perturbation in M_s at 20 seconds.) While there actually were slight changes in the $m_b^{(1)}$ values for the two cases, they were too small to be shown on this plot. However, with stress concentrations present they could be considerably larger.

Also shown on the figure are lines indicating earthquake populations for different regions, and points (x) denoting some of the more anomalous NTS explosions (generally those having high M_s values relative to their m_b values). Note that some of these events fall close to the extreme values for M_s perturbations due to this kind of tectonic release. Comparing them with the earthquake population limit lines shows that they could be confused with earthquakes on the basis of the m_b-M_s discrimination criteria alone. Finally, for general reference, the mean line for explosions at NTS is also shown.

Summary: State of Knowledge and Research Needs

The previous discussion is, in effect, a summary of what is considered to be the current state of knowledge. In addition, it is to be hoped that the uncertainties and ambiguities are reasonably well covered. By way of a summary, then, it seems most useful to briefly state the essential conclusions to be drawn, albeit with some being rather tenative, and to then list areas of research that could provide the necessary details for very firm conclusions.

The principal conclusions and results are:

(1) Stress relaxation effects can be expected in any material capable of sustaining long term non-hydrostatic stress. The largest effects will occur in the (known) regions of high tectonic activity. That is, large effects would be expected at plate margins in and near intrusive zones and generally where loading of the crust is evident, such as near large river deltas. Because of the likelihood of high stresses in seismically quiet zones along plate boundaries, it is probable that seismic gaps are areas of special importance.

(2) It appears that most of the well documented anomalous effects in the seismic radiation field from explosions (large SH wave production, observed especially at long periods from 10-25 seconds; strong perturbations in the long period Rayleigh type surface waves, azimuthally dependent increase in complexity of the short period P waves) are due to tectonic stress relaxation in the vicinity of

the roughly spherical shatter zone created by the explosive shock wave. However, there may be, at least in some cases, strong asymmetries in the fracture zone around an explosion; especially when the medium is highly stressed and/or existing stress loaded faults are nearby. In this case triggering of an earthquake is said to have occurred. However, the liklihood of this occurrence is judged to be small and to have rarely occurred in the past on a large scale. Nevertheless, it is possible that future tests could be so arranged so as to maximize the likelihood of such an occurrence. When an earthquake at or very near the test point is induced, discrimination and especially yield estimation would be considerably more difficult.

(3) Long period surface wave radiation is strongly perturbed by tectonic release effects within the whole measureable low frequency band (i.e., from approximately 5 sec. to 60 sec. in period). The perturbations in the observed Rayleigh wave forms can be such as to add to, or subtract from, the explosive generated Rayleigh wave depending on the orientation and magnitude of the prestress in the vicinity of the explosion. The magnitude of this effect can be very (unacceptably) large. In those cases where Love waves are significant, so that tectonic release is involved, then yield estimation using M_s can only be made after correction of the Rayleigh wave measurement using the observed Love wave to deduce the size and configuration of the tectonic source. Such a correction would be much more reliable, when spherical shatter zone induced tectonic release is involved, than it would be if actual earthquake triggering is involved.

(4) Short period perturbations in the wave train can be expected to be very complex due to dependence on stress concentration effects and local complex structure. However, the perturbations should be small to moderate for the first cycle of the P wave motion, while being significantly larger for the later part of the P wave train. Body wave magnitude measured from the first P wave cycle should, therefore, be minimally perturbed by stress relaxation effects and the complexities of local structure. Further, corrections to the first cycle of the P wave for tectonic affects could conceivably be made for purposes of yield estimation.

The current state of understanding of the importance and precise nature of tectonic release effects could be considerably expanded if research were initiated to:

(1) Pursue detailed studies of selected events involving moderate to large anomalous contributions with the objective of fitting both body and surface waves simultaneously in order to determine the anomalous source component as completely and accurately as possible.

(2) Investigate the correlation between explosion event anomalous characteristics and regional tectonics for both the U.S. and U.S.S.R.

(3) Systematically investigate the nature and size of the P wave perturbations for non-uniform initial stress (including lithostatic stress alone) using the available analytical model for shatter zone induced stress release effects.

(4) Study observational analysis methods designed to minimize P wave magnitude perturbations from stress release effects. Perfect methods using synthetic seismograms and then apply to explosion events with known yields as a check.

(5) Predict M_s perturbations possible for various realistic prestress environments. Consider methods for minimization of perturbations, for both yield and discrimination purposes.

(6) Consider methods, such as that which could be based on inferrence of the tectonic component of the composite source using Love and short period S waves, to correct M_s and m_s measurements for tectonic component effects.

(7) Initiate 3-D numerical modeling studies of explosions in prestressed media; to obtain estimates of shatter zone size and symmetry, as well as to predict the P and S waves from the source, all as functions of material type, prestress levels and orientation, etc.

REFERENCES

Aki, K. and Y. B. Tsai (1972), "Mechanism of Love-Wave Excitation by Explosive Sources," JGR, 27, pp. 1452-1475, March.

Archambeau, C. and C. Sammis (1970), "Seismic Radiation from Explosions in Prestressed Media and the Measurement of Tectonic Stress in the Earth," Reviews of Geophysics and Space Physics, 8, August.

Archambeau, C. B. (1972), "The Theory of Stress Wave Radiation from Explosions in Prestressed Media," Geophys. J. R. Astr. Soc., 29, pp. 329-366.

Archambeau, C. B., D. G. Harkrider and D. V. Helmberger (1974), "Studies of Multiple Seismic Events," ACDA Final Report, Contract ACDA/ST-220.

Bache, T. C. (1976a), "The Effect of Tectonic Stress Release on Explosion P-Wave Signatures," BSSA, 66, pp. 1441-1457, October.

Bache, T. C. (1976b), "The Seismological Evidence for the Triggering of Block Motion by Large Explosions," DNA Report, Contract No. DNA001-76-C-0072, December.

Kisslinger, C. (1976), "A Review of the Mechanisms of Induced Seismicity," Eng. Geol., 10, pp. 85-98.

Lambert, D. G., E. A. Flinn and C. B. Archambeau (1972), "A Comparative Study of the Elastic Wave Radiation from Earthquakes and Underground Explosions," Geophys. J. R. Astr. Soc., 29, pp. 403-432.

Stevens, J.L. (1980), "Seismic Radiation from the Sudden Creation of a Spherical Cavity in an Arbitrarily Prestressed Elastic Medium," Geophys. J. Roy. Astron. Soc., in press.

Toksoz, M.N., D.G. Harkrider and A. Ben-Menahem (1965), "Determination of Source Parameters by Amplitude Equalization of Seismic Surface Waves: Part 2. Release of Tectonic Strain by Underground Nuclear Explosions and Mechanisms of Earthquakes," J. Geophys. Res., 70, pp. 907-922.

Toksoz, M. N. and H. H. Kehrer (1972), "Tectonic Strain Release by Underground Nuclear Explosions and Its Effect on Seismic Discrimination," Geophys. J. R. Astr. Soc., 31, pp. 141-161.

11. TECTONIC GENERATION

by

Thomas C. Bache

Systems, Science and Software
P. O. Box 1620
La Jolla, California 92038

Introduction

Almost from the beginning of the underground nuclear test program in Nevada it was noticed that the seismic radiation included earthquake-like features. This, together with geological evidence, has led to the hypothesis that explosions release a substantial amount of tectonic strain energy. The literature on this subject is extensive. For example, Aki and Tsai (1972) summarize much of the evidence for concluding that the explosions trigger earthquake faulting. We reviewed this subject some years ago (Bache and Lambert, 1976) and found some serious weaknesses in the argument that explosions commonly trigger moderate-sized earthquakes. The aftershock distributions, where mapped (e.g., Hamilton ahd Healey, 1969), seem to suggest that the tectonic stress is released in a more localized region about the explosion. Thus, we tended to favor a mechanism more closely related to that proposed by Archambeau (1972) in which stored tectonic stresses are released within a volume of seriously disturbed material about the cavity.

In either case, the conventional wisdom is that superimposed on the explosion source is a double-couple aligned with the local stress field. From a theoretical point-of-view, one should be able to deduce the orientation and moment of this double-couple from the radiated Love and Rayleigh waves and correct the data for this effect. Toksöz and Kehrer (1972) did this for a substantial number of events. However, they introduced the restrictive assumptions that the double-couple

sour had the same location and time function as the explosion and ˎ ꞩ oriented as a vertical strike-slip fault. If enough good data were available, these assumptions could be relaxed. For example, Rivers and von Seggern have been inverting for the best moment tensor fit to the PILEDRIVER data using the moment tensor inversion scheme of Mendiguren (1977). Of course, any inversion scheme is limited by the accuracy of the path corrections used.

How well does the explosion plus double-couple model fit the data? The strike-slip double-couple used in the inversions of Toksöz and Kehrer does not give a particularly good fit. Rivers and von Seggern (1978) seem to fit the PILEDRIVER data quite well, but with a model for which the physical interpretation is not entirely clear. The orientation is that of a reverse fault and both the explosion and double-couple are at a depth below 2.5 km in an earth model that did not include the Climax Stock granite.

It does seem clear that the explosion plus double-couple model qualitatively explains the long period data. Perhaps it also gives a quantitative explanation that can be used to correct individual station amplitudes. But more detailed work is necessary to prove this concept.

The contribution of a double-couple can certainly be obscured by other effects. For example, under Topic 13 we summarize a very deta ed study of Rayleigh wave amplitudes at ALQ and TUC. After c ecting for the path, the amplitudes at TUC are systematicall rger than those at ALQ. The mean ratio is about 1.5. T. ratio seems too consistent from event-to-event to be due enti ' to tectonic strain release contributions.

Double-Couple Contribution to Short Period Surface Waves

Our major interest is correcting for a double-couple contribution to the surface waves at long periods. While the

simple model may work fairly well at these periods, there are shorter period seismic observations that are not explained very well. This gives rise to some concern that there are important features of the physical process that are poorly understood.

First, consider SALMON. Archambeau, Flinn and Lambert (1966) point out that this event showed no evidence of a double-couple component at long periods. However, the short period Lg transverse motion is about the same size as the other two components. Smart (1978) shows that the transverse motion is orthogonal to the radial Lg along a path extending back through the source. This may be an indication that the transverse motion results from wave conversion along the path, or it may indicate that the SALMON source generated substantial SH motions at high frequencies.

Analysis of Regional Motions from MIGHTY EPIC and DIABLO HAWK

Bache, Farrell and Lambert (1979) describe an analysis of the regional ground motions of the Rainier Mesa explosions DIABLO HAWK and MIGHTY EPIC. These were very similar events of fairly low yield. The recording stations, seven for MIGHTY EPIC and ten for DIABLO HAWK, were broad-band, three-component digitally recorded stations operated by Lawrence Livermore Laboratory, Sandia Laboratories and Systems, Science and Software (S^3). The distances were 131 to 368 km.

The objective of the experiment was to infer the size and orientation of the double-couple component associated with these events. First, the data were processed to determine the following:

- The coherence between the same station recordings of the two events is very high. The sources were, therefore, identical except for a scale factor which may be azimuthally dependent.

- The coherence between different components is low. This seems to indicate that the horizontal component waves emanate from the vicinity of the source.

- The data were filtered with a four to eight second band pass filter and with MARS to identify the fundamental mode Love and Rayleigh waves with a period near six seconds. All indications are that these modes were successfully isolated.

The best fitting explosion RDP plus double-couple was determined by fitting the LR, LQ and LQ/LR data. The double-couple was assumed to be a vertical strike-slip. The best fitting solutions for DIABLO HAWK are shown in Figures 1, 2 and 3. The analogous solutions for MIGHTY EPIC are shown in Figures 4 and 5.

The inferred RDP, (Ψ_∞) is within the range of expected values for such events. This indicates that the amplification effects of the path are about right at these regional distances. The fit to the Love waves is not very good. In fact, the data would be fit nearly as well with a circle!

There is another interesting and puzzling aspect of these data. The DIABLO HAWK/MIGHTY EPIC ratio is about 0.9 at high frequencies and 0.6 at low frequencies. This is true for nearly all components at all stations. The frequency domain ratios can also be seen in time domain measurements of P_n (ratio of 0.9) and long period Love and Rayleigh waves (ratio of 0.6). It seems that the total, explosion plus double-couple, source level is larger at long periods for DIABLO HAWK. There is no easy way to explain this in the context of the simple theory used.

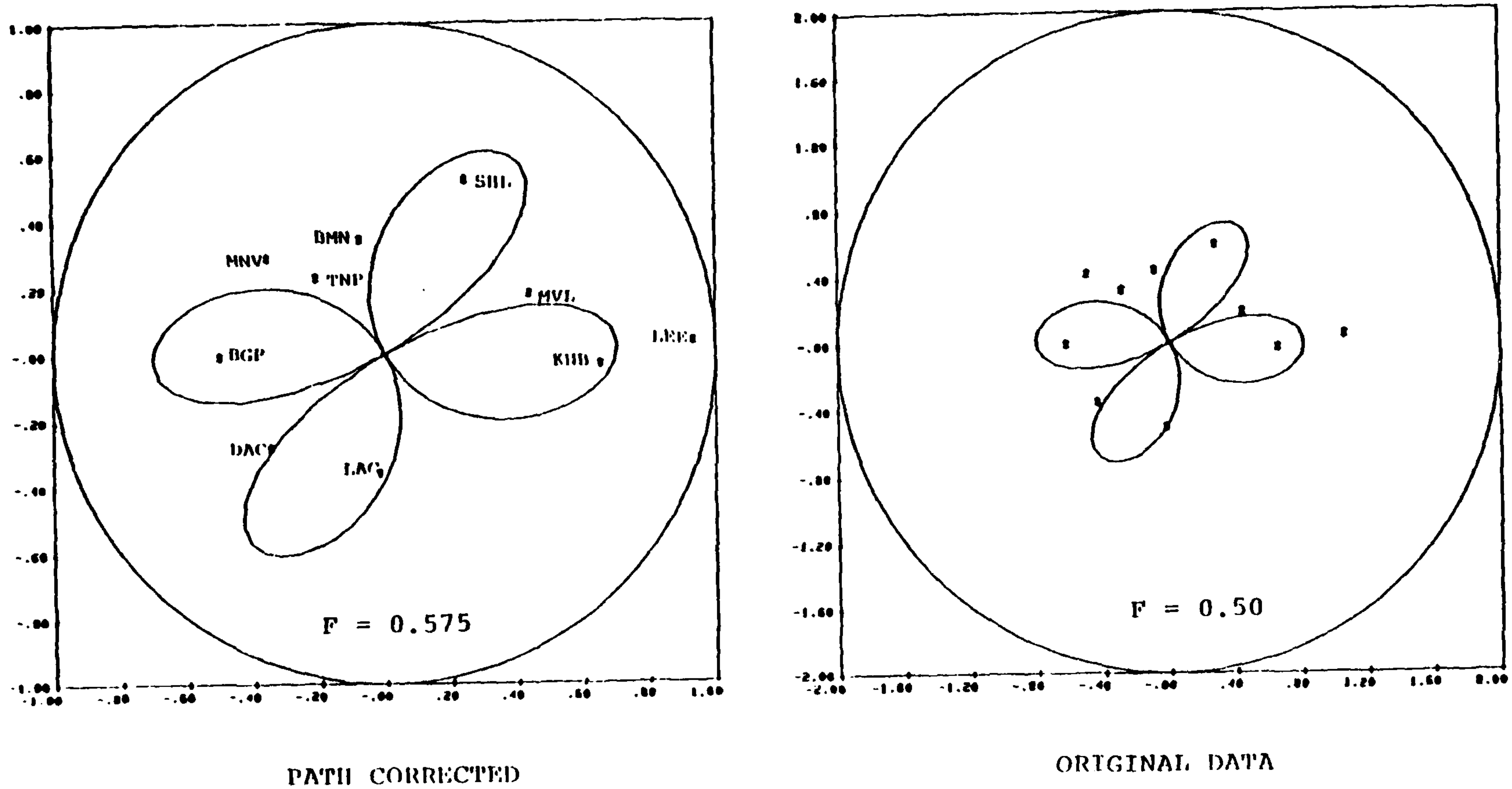


Figure 1. Comparison of theoretical LQ/LR radiation patterns (δ = 90, λ = 0, strike = N16°E) to the observed data with and without path corrections.

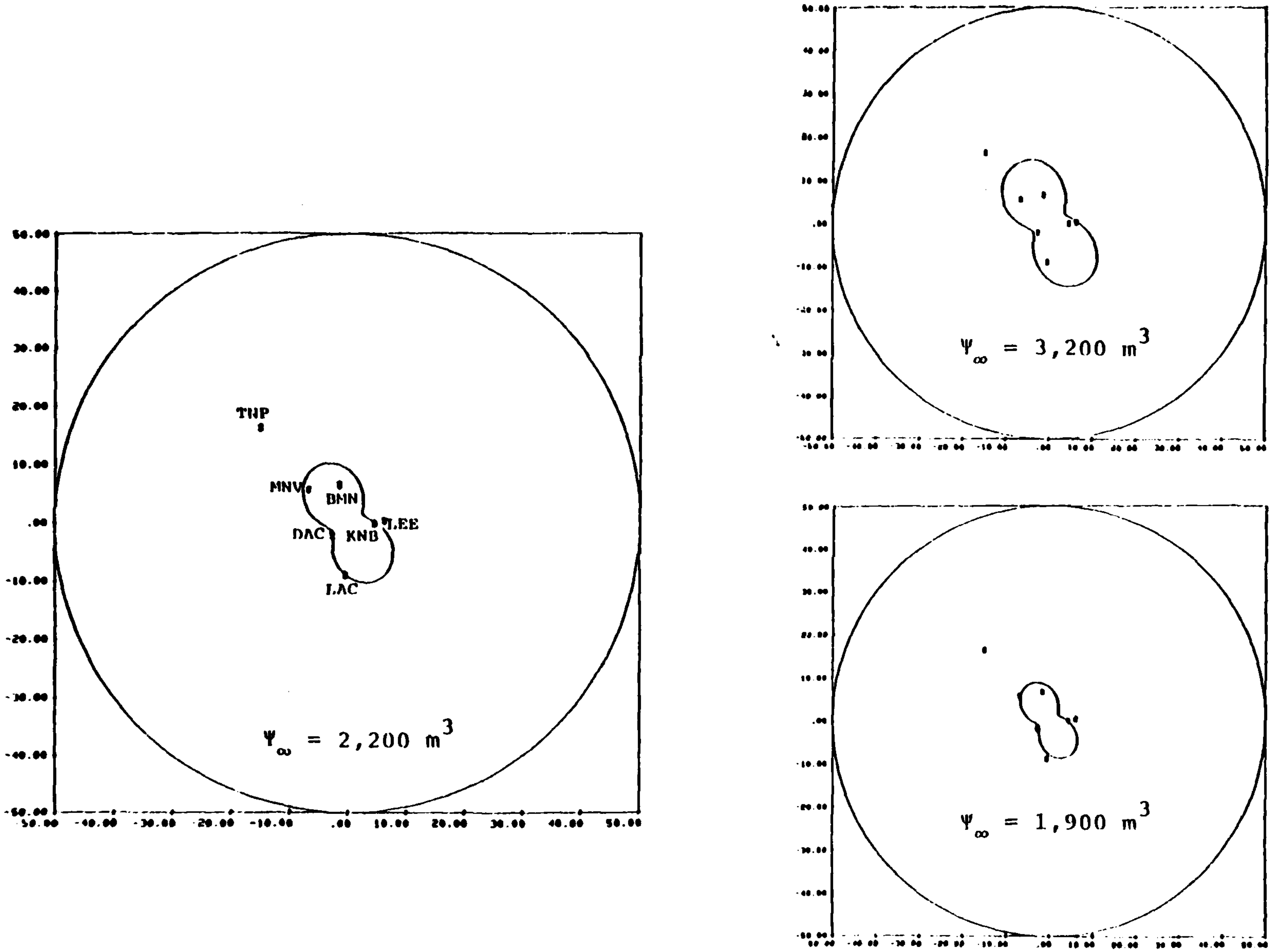


Figure 2. Theoretical Rayleigh wave radiation patterns are compared to the DIABLO HAWK observations normalized to a range of 250 km. The theoretical solution has $\delta = 90$, $\lambda = 0$, strike = N16°E, $F = 0.575$, $\mu_s = 40$ kbar and three values of Ψ_∞.

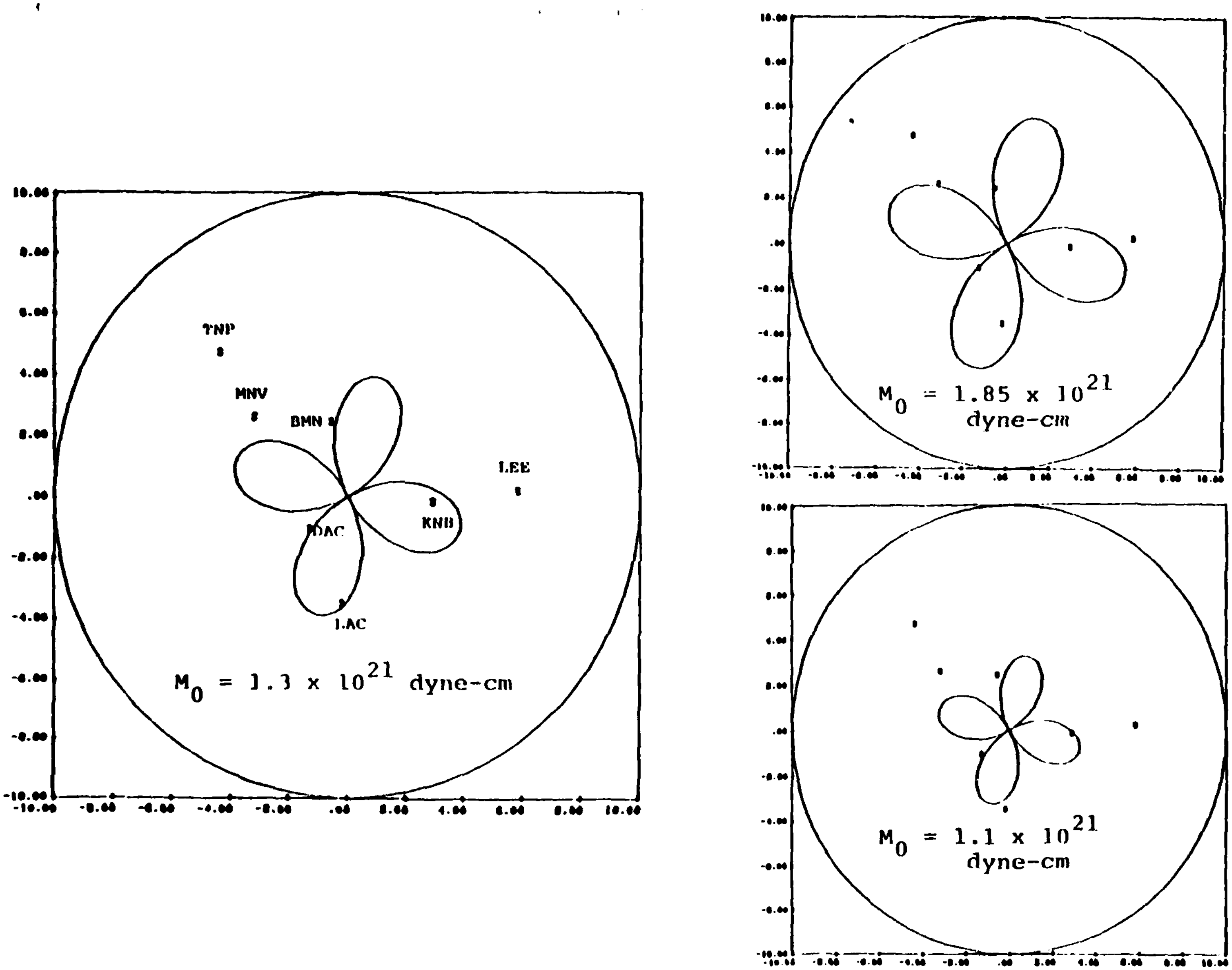


Figure 3. Theoretical Love wave radiation patterns are compared to DIABLO HAWK observations normalized to 250 km. The theoretical solutions have $\delta = 90$, $\lambda = 0$, strike = N16°E, F = 0.575, μ_s = 40 kbar and three values of the moment, M_0.

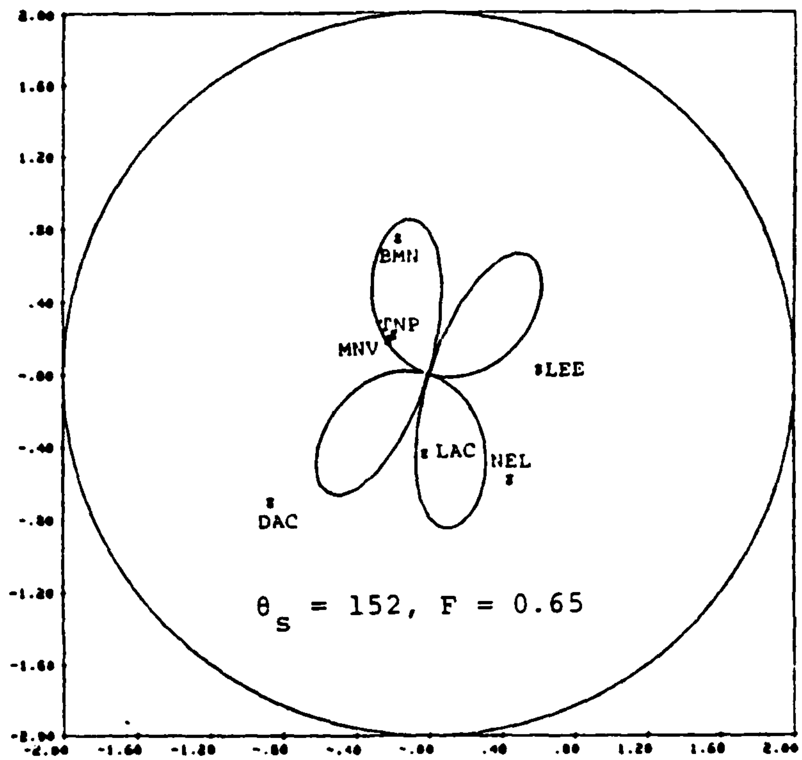


Figure 4. The fit to the MIGHTY EPIC LQ/LR data. We assume vertical strike-slip faulting.

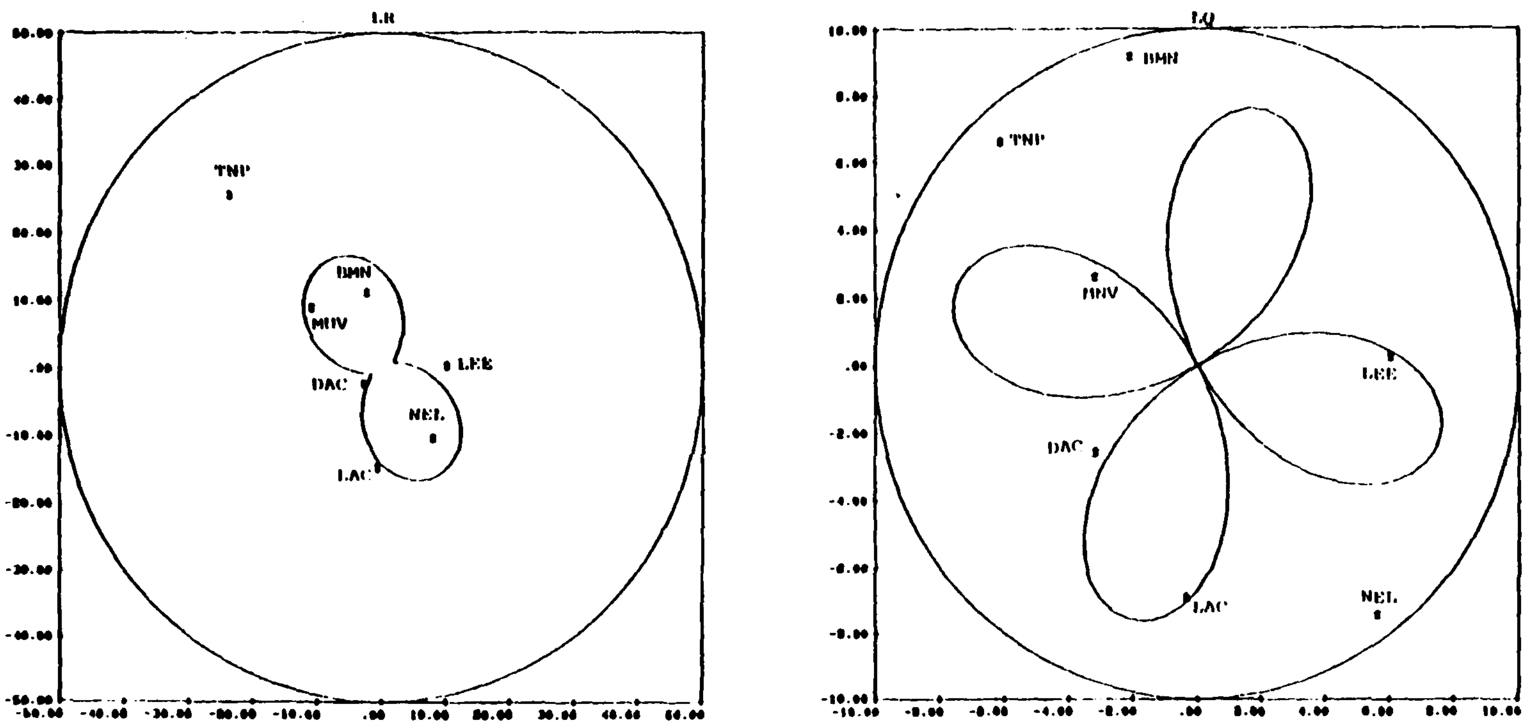


Figure 5. MIGHTY EPIC LR and LQ data compared to theoretical radiation patterns for $\theta_s = 16$, $\delta = 90$, $\lambda = 0$, $\Psi_\infty = 3{,}200\ m^3$, $F = 0.8$, $M_0 = 2.6 \times 10^{21}$ dyne-cm.

Conclusions

Considering nuclear explosions to be an explosion plus a double-couple generated by some kind of tectonic strain release gives a reasonable qualitative explanation of observed long period seismic waves. The double-couple causes a Rayleigh wave radiation pattern. The size of this effect compared to amplitude fluctuations due to path variations has not been conclusively demonstrated to my knowledge.

The mechanism for generating the double-couple is not very well understood. Triggered fault motion and volume source release of tectonic strain energy must both occur to some degree. Passive slippage along planes of weakness is also known to occur and this can lead to substantial amounts of wave conversion (e.g., Salvado and Minster, 1979).

At high frequencies, for example, those important for Lg, the mechanism for generating transverse motion is not well understood. Our study of MIGHTY EPIC and DIABLO HAWK was at fairly long periods, six to eight seconds. There were good sized Love waves for these events, but they are not fit very well by a double-couple radiation pattern. Perhaps the path applied corrections substantially bias the data. More likely, the Love wave source is not a simple double-couple, but has a more uniform radiation pattern. The amplitude ratios between these two very similar events are also difficult to explain with conventional explosion plus double-couple models.

REFERENCES

Aki, K. and Y. Tsai (1972), "The Mechanics of Love Wave Excitation by Explosive Sources," JGR, 77. pp. 1452-1475.

Archambeau, C. B., E. A. Flinn and D. G. Lambert (1966), "Detection, Analysis and Interpretation of Teleseismic Signals: 1. Compressional Phases from the SALMON Event," JGR, 71, pp. 3483-3501.

Archambeau, C. B. (1972), "The Theory of Stress Radiation from Explosions in Prestressed Media," Geophys. J. R. Astr. Soc., 29, pp. 329-366.

Bache, T. C., W. E. Farrell and D. G. Lambert (1979), "Block Motion Estimates from Seismological Observations of MIGHTY EPIC and DIABLO HAWK," Systems, Science and Software Final Report SSS-R-79-4080, submitted to the Defense Nuclear Agency, July, 176 pages.

Bache, T. C. and D. G. Lambert (1976), "The Seismological Evidence for the Triggering of Block Motion by Large Explosions," Systems, Science and Software Report, DNA 4323T.

Hamilton, R. M. and J. H. Healey (1969), "Aftershocks of the BENHAM Nuclear Explosion," BSSA, 59, pp. 2271-2281.

Mendiguren, J. A. (1977), "Inversion of Surface Wave Data in Source Mechanism Studies," JGR, 82, pp. 889-894.

Rivers, W. and D. H. von Seggern (1978), "Release of Tectonic Strain Energy by PILEDRIVER," unpublished Teledyne Geotech memorandum, September.

Salvado, C. and J. B. Minster (1979), "Slipping Interfaces: A Possible Source of S Radiation from Explosive Sources," submitted to BSSA for publication.

Smart, E. (1978), "A Three-Component, Single-Station, Maximum-Liklihood Surface Wave Processor," Teledyne Geotech Report SDAC-TR-77-14, submitted to AFTAC/VSC, July.

Toksöz, M. N. and H. H. Kehrer (1972), "Tectonic Strain Release by Underground Nuclear Explosions and Its Effect on Seismic Discrimination," Geophys. J., 31, pp. 141-161.

California Institute of Technology, Pasadena

Tectonic Generation

David G. Harkrider

A lot of work has been done on the effect of superimposing a vertical pure strike slip double couple like mechanism on explosion generated surface waves. This tectonic generation model was felt to be most appropriate to NTS and the French Sahara Test site (Harkrider (1977) and Rodi *et al*. (1978).

Another and potentially more troublesome tectonic mechanism is the 45° dipping pure thrust fault or tectonic release due to elastic failure in an equivalent orientation of a pre-stress field. If strong enough the additional seismic radiation can reverse the Rayleigh wave signature which would be observed for the explosion alone.

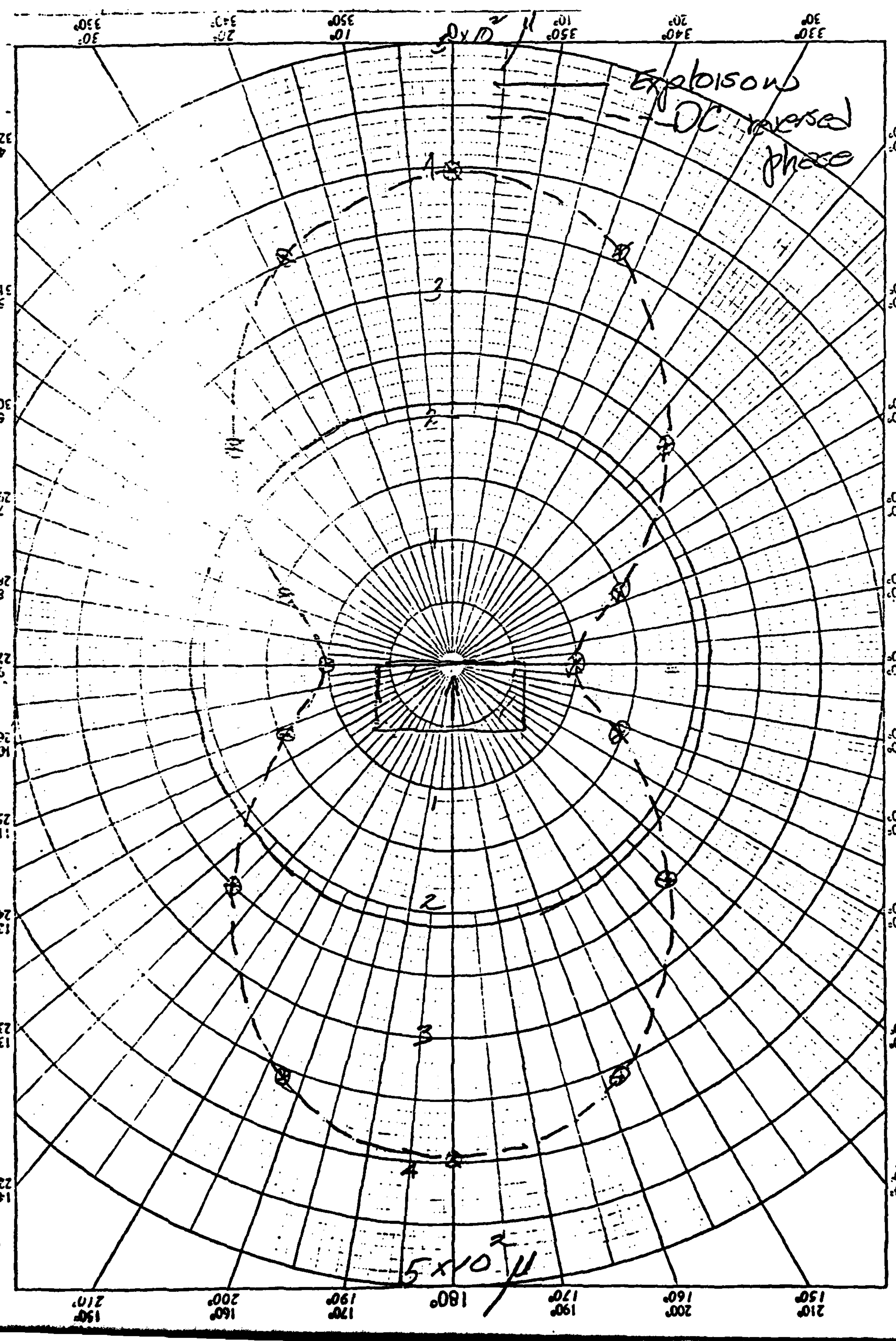
Attached are synthetic fundamental Rayleigh waves for an explosion and a pure thrust 45° dipping double couple at a range of 4000 km and enough azimuths to determine the radiation pattern for the double couple. The two mechanisms are synthesized separately and the phase of the double couple is found to be reversed with respect to the explosion at all azimuths. The peak airy phase amplitude and log amplitude are plotted as a function of azimuth. The polar figures are scaled such that the P wave moment for both the explosion and the double couple are the same, $M_o = 10^{25}$ (dyne-cm). This does not imply that their P waves have the same amplitude since this is strong function of source rise time and was not investigated here.

References

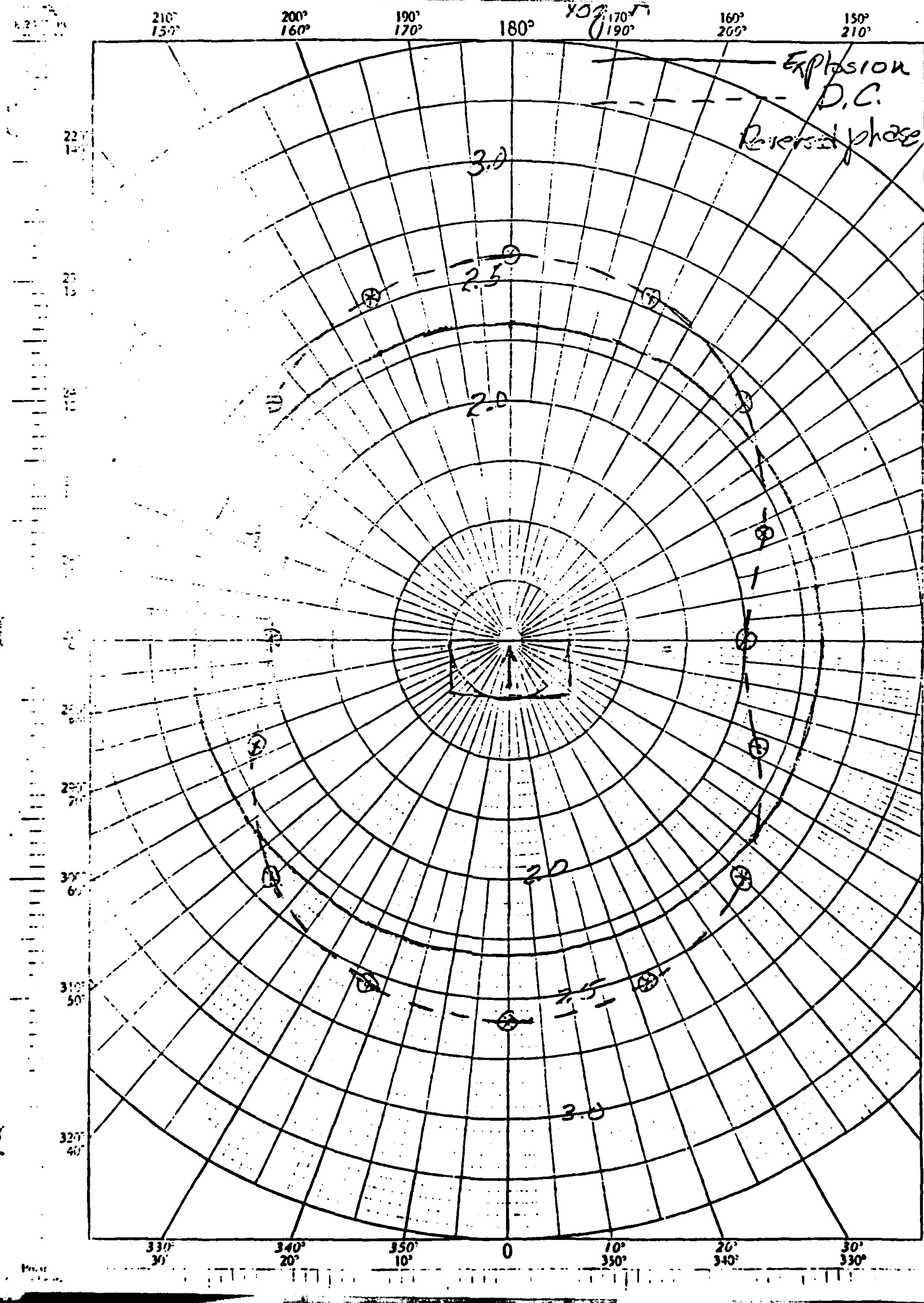
Harkrider, David G. (1977), Preliminary Report on the French Sahara Event Saphire, Semi-Annual Technical Report 1 April 1977-30 September 1977, California Institute of Technology, AFOSR #F49620-77-C-0022.

Rodi, W. L., J. M. Savino, T. G. Barker, S. M. Day, and T. C. Bache (1978), Analyses of Explosion Generated Surface Waves in Africa, Results from Discrimination Experiment and Summary of Current Research, Systems, Science and Software Quarterly Technical Report SSS-R-78-3653.

Best
Available
Copy



Explosion
DC reversed phase

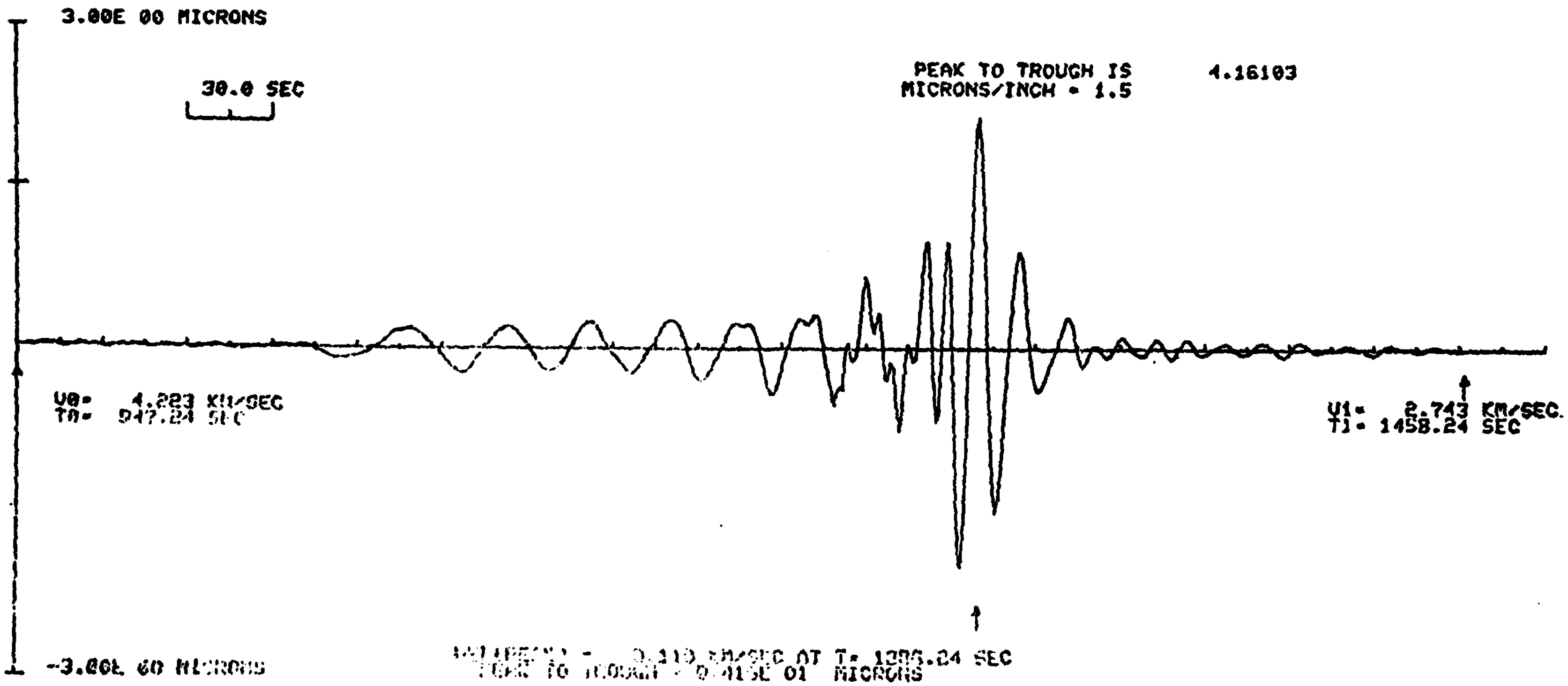


210° 150°
200° 160°
190° 170°
180°
170° 190°
160° 200°
150° 210°
Explosion
D.C.
Reversed phase
3.0
2.5
2.0
140° 220°
130° 230°
120° 240°
110° 250°
100° 260°
90° 270°
80° 280°
70° 290°
60° 300°
50° 310°
40° 320°
2.0
2.5
3.0
330° 30°
340° 20°
350° 10°
0
10° 350°
20° 340°
30° 330°

```
RESET SEISMOGRAM SCALES ? (Y OR N) : Y
CHANGE TIME SCALE ? (Y OR N) : N
CHANGE VERTICAL SCALE OPTION ? (Y OR N) : Y
SET SEISMOGRAM VERTICAL SCALE ? (Y OR N) : Y
```

$$M_o = 4\pi\rho\alpha^2\Psi(\infty) = 1.98\times10^{23} \text{ (dyne-cm)}$$

```
                    EXPLOSION ONLY
                    SOURCE DEPTH =      0.300000
ALPS =  6.2000 ,BETS =  3.5110 ,RHOS =  2.7360 ,MUS = 33.727
                    DISTANCE =     4000.00
GRANITE.328.RVP!
YIELD =  135.0 KT, ZI(INF) = 0.1498E 05  m^3

CIT 109  RAYLEIGH FUNDAMENTAL (WWSN)
```



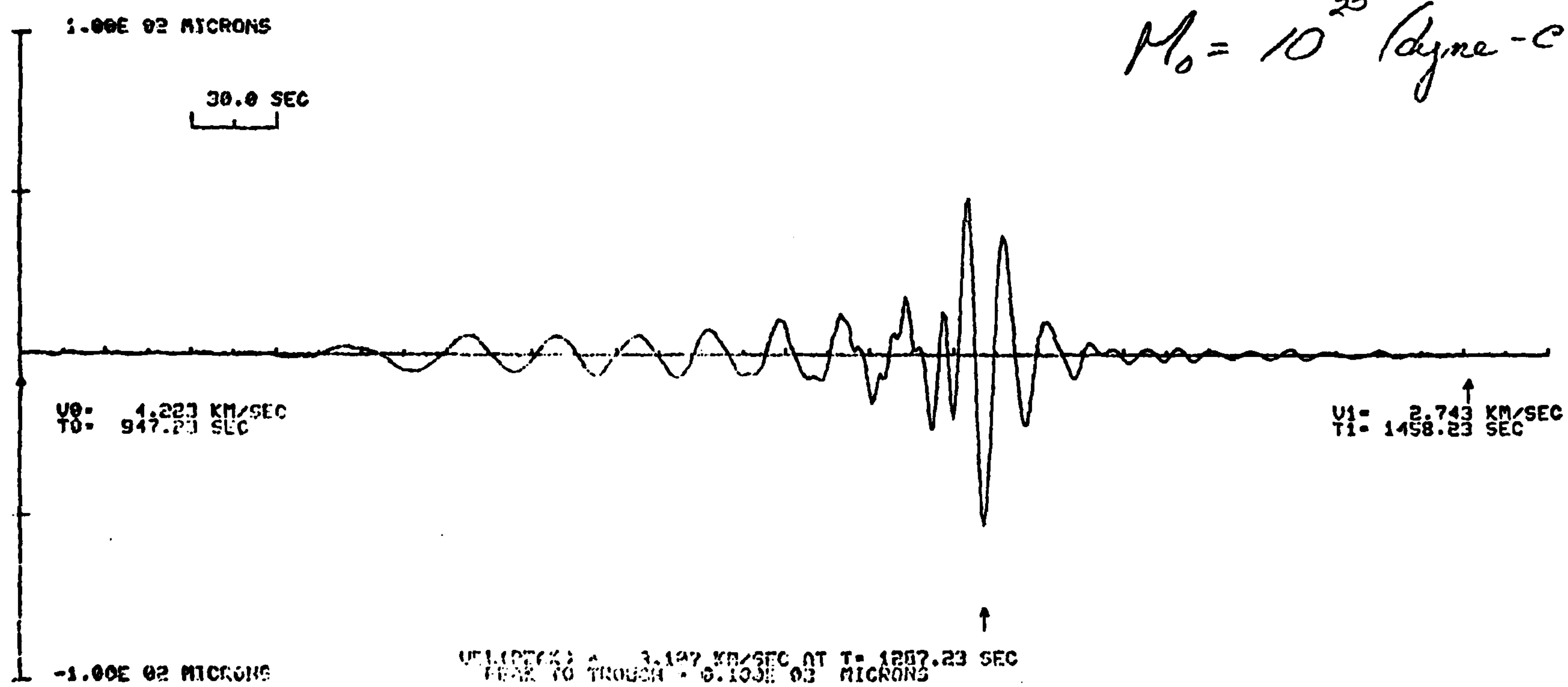
SAVE SEISMOGRAM ? (Y OR N) :

PAUSE

```
DISTANCE =    4000.00      AZIMUTH =       0.00
INCLUDE EXTERNAL EXPLOSION ? (Y OR N) : N
ONE SIDED OR DEGRADED SOURCE POTENTIALS ? (Y OR N) : N
ATTENUATION ? (Y OR N) : Y
CHANGE ATTENUATION ? (Y OR N) : N
INSTRUMENT ? (Y OR N) : Y
HIGHER MODE ? (Y OR N) : N
INITIAL OR RESETTING INITIAL SEISMOGRAM PARAMETERS ? (Y OR N) : N
LFFT =      3 NT =      512
INITIAL OR RESETTING INITIAL SEISMOGRAM SCALES ? (Y OR N) : N
```

```
        POINT DOUBLE COUPLE OR DISLOCATION
        SOURCE DEPTH =      0.300000
ALPS =  6.2000 ,BETS =  3.5110 ,RHOS =  2.7360 ,MUS = 33.7276
        DELTA = 45.00 ,LAMBDA = 90.00
        DISTANCE =      4000.00      AZIMUTH =      0.000000
BRUNE FAR FIELD BODY WAVE RISE TIME =   0.5000 SECS

CIT 109  RAYLEIGH FUNDAMENTAL (WWSN)
```

$$M_0 = 10^{25}\ (\text{dyne-cm})$$

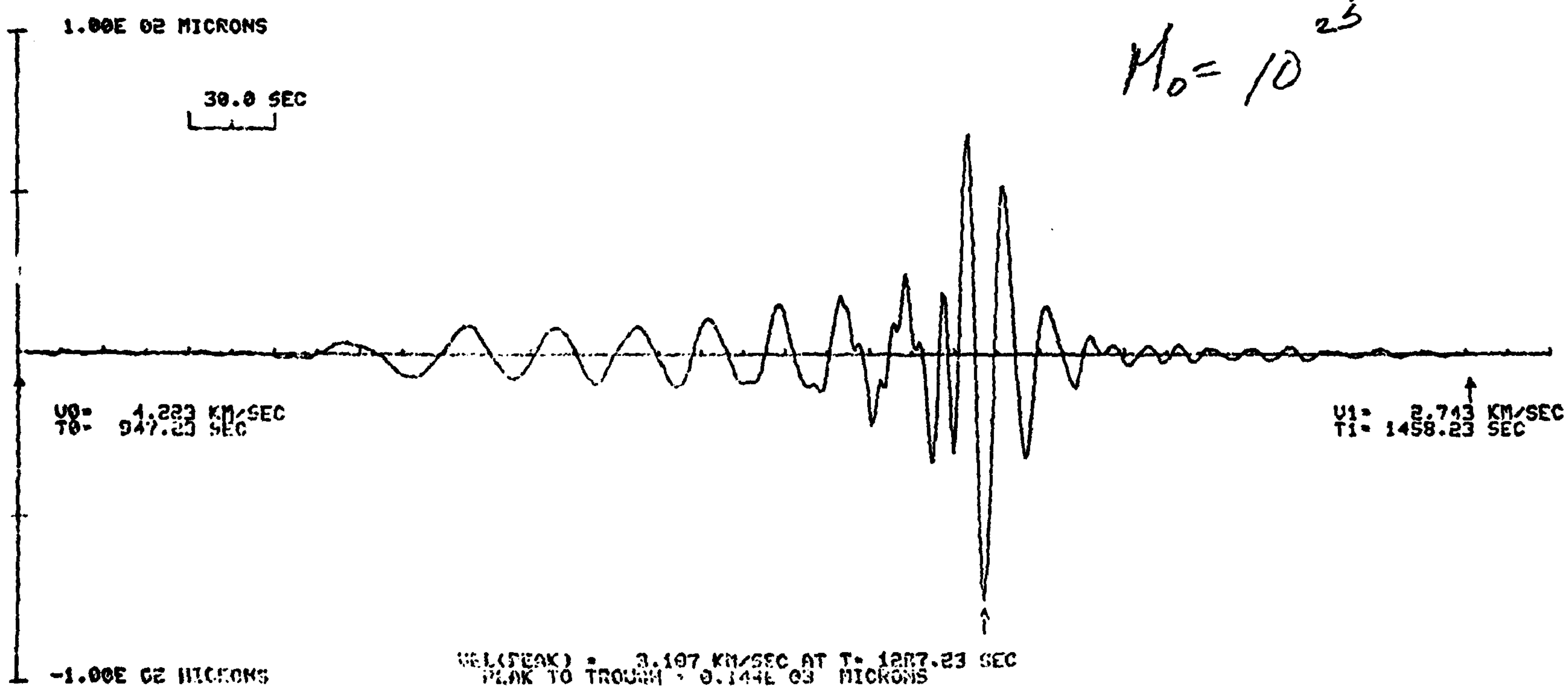


SAVE SEISMOGRAM ? (Y OR N) : N

```
RESET SEISMOGRAM SCALES ? (Y OR N) : N
CHANGE SEISMOGRAM PARAMETERS ? (Y OR N) : N
CHANGES TO THIS SEISMOGRAM ?
  CIT 109  RAYLEIGH FUNDAMENTAL (UWSN)
        (Y OR N) : Y
CHANGE INSTRUMENT OR ATTENUATION OPTIONS ? (Y OR N) : N
MODIFY SOURCE ? (Y OR N) : N
A NEW DISTANCE OR AZIMUTH ? (Y OR N) : Y
DISTANCE, AZIMUTH :4000,22.5
 DISTANCE =    4000.00      AZIMUTH =     22.50
INCLUDE EXTERNAL EXPLOSION ? (Y OR N) : N
ONE SIDED OR DEGRADED SOURCE POTENTIALS ? (Y OR N) : N
ATTENUATION ? (Y OR N) : Y
CHANGE ATTENUATION ? (Y OR N) : N
INSTRUMENT ? (Y OR N) : Y
HIGHER MODE ? (Y OR N) : N
INITIAL OR RESETTING INITIAL SEISMOGRAM PARAMETERS ? (Y OR N) : N
LFFT =     3 NT =     512
INITIAL OR RESETTING INITIAL SEISMOGRAM SCALES ? (Y OR N) : N
```

```
      POINT DOUBLE COUPLE OR DISLOCATION
      SOURCE DEPTH =      0.300000
ALPS =  6.2000 ,BETS =  3.5110 ,RHOS =  2.7360 ,MUS = 33.727
      DELTA = 45.00 ,LAMBDA = 90.00
      DISTANCE =      4000.00      AZIMUTH =      22.5000
BRUNE FAR FIELD BODY WAVE RISE TIME =   0.5000 SECS

 CIT 109  RAYLEIGH FUNDAMENTAL (UWSN)
```

$M_0 = 10^{25}$

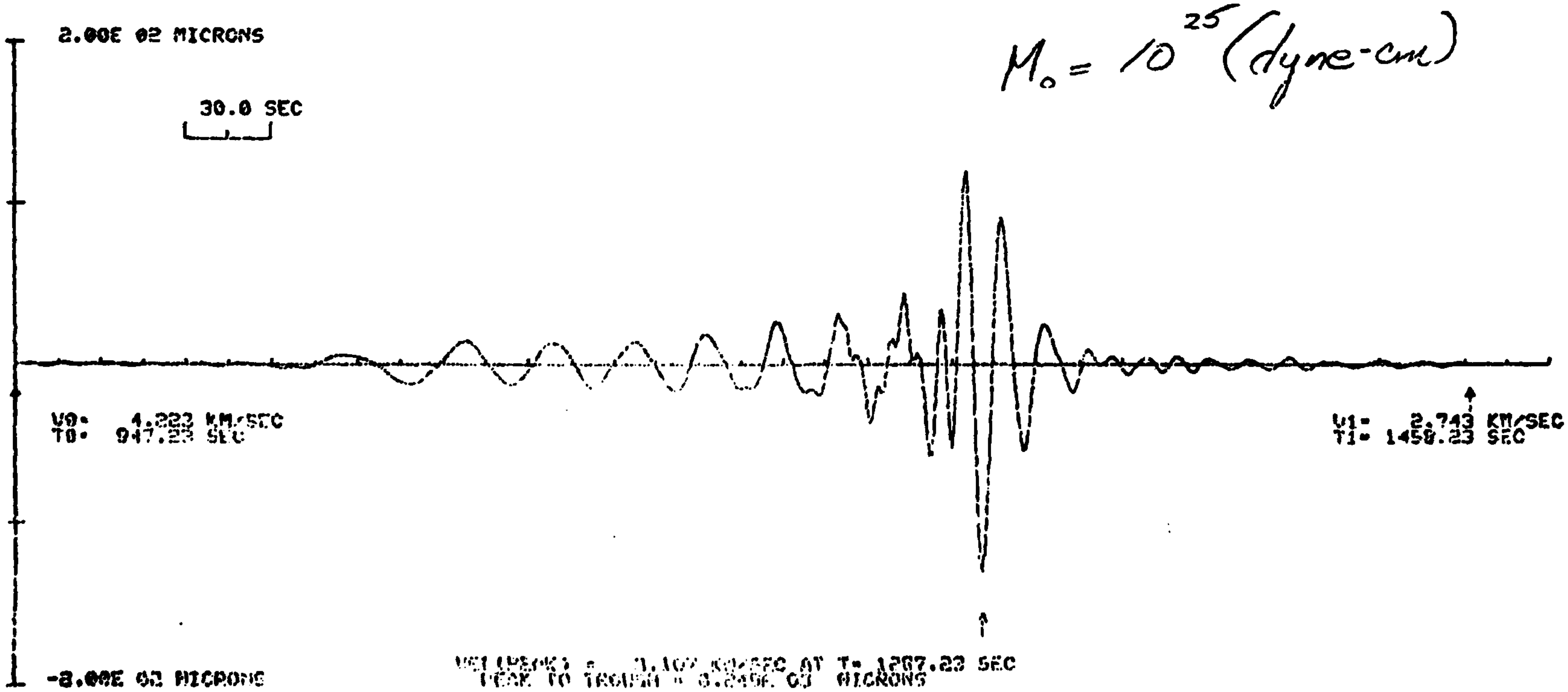



```
SAVE SEISMOGRAM ? (Y OR N) :
```

```
4000,45
 DISTANCE =   4000.00      AZIMUTH =      45.00
INCLUDE EXTERNAL EXPLOSION ? (Y OR N) : N
ONE SIDED OR DEGRADED SOURCE POTENTIALS ? (Y OR N) : N
ATTENUATION ? (Y OR N) : Y
CHANGE ATTENUATION ? (Y OR N) : N
INSTRUMENT ? (Y OR N) : Y
HIGHER MODE ? (Y OR N) : N
INITIAL OR RESETTING INITIAL SEISMOGRAM PARAMETERS ? (Y OR N) : N
LFFT =      3 NT =     512
INITIAL OR RESETTING INITIAL SEISMOGRAM SCALES ? (Y OR N) : N
```

```
           POINT DOUBLE COUPLE OR DISLOCATION
           SOURCE DEPTH =      0.300000
ALPS =  6.2000 ,BETS =  3.5110 ,RHOS =  2.7360 ,MUS = 33.727
           DELTA = 45.00 ,LAMBDA = 90.00
           DISTANCE =      4000.00     AZIMUTH =      45.0000
BRUNE FAR FIELD BODY WAVE RISE TIME =   0.5000 SECS

 CIT 109  RAYLEIGH FUNDAMENTAL (WWSN)
```

$M_o = 10^{25}$ (dyne-cm)

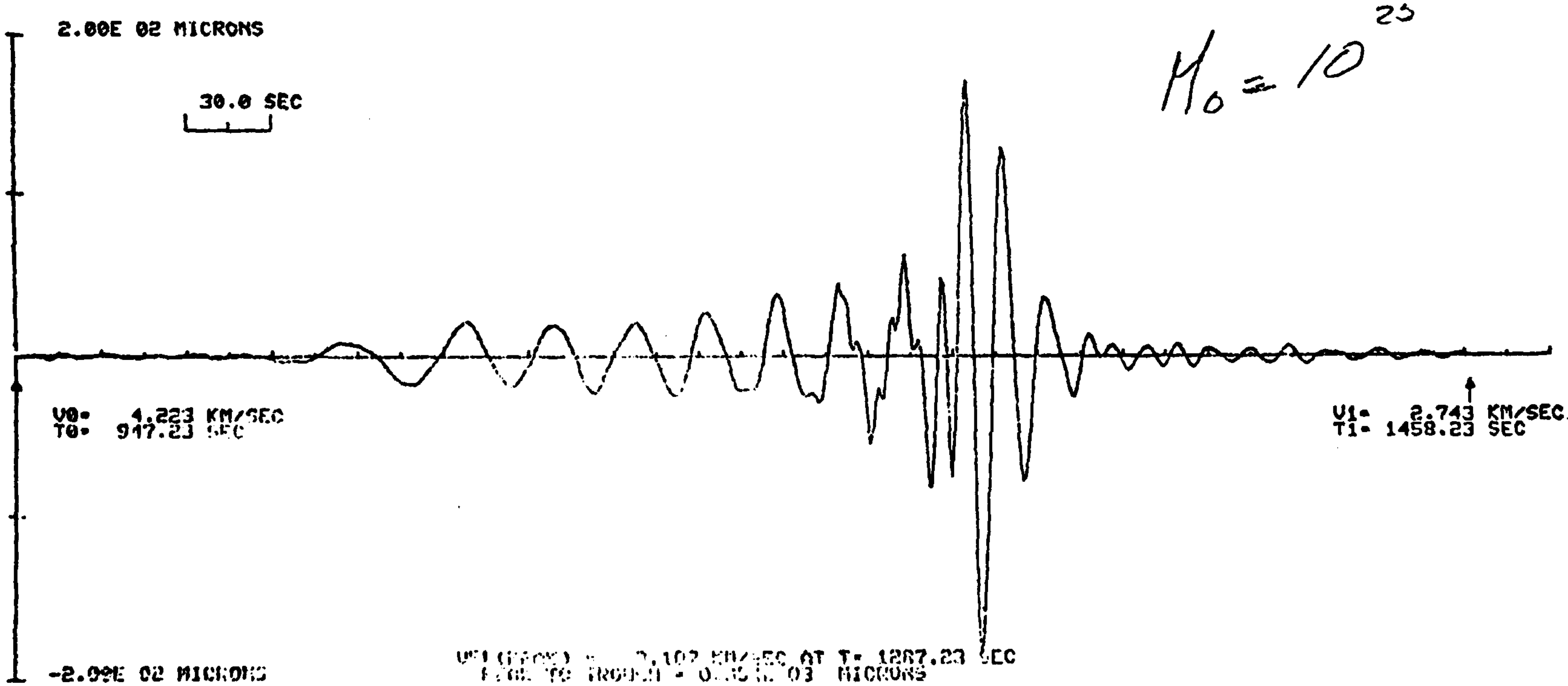



```
SAVE SEISMOGRAM ? (Y OR N) :
```

```
RESET SEISMOGRAM SCALES ? (Y OR N) : N
CHANGE SEISMOGRAM PARAMETERS ? (Y OR N) : N
CHANGES TO THIS SEISMOGRAM ?
  CIT 109  RAYLEIGH FUNDAMENTAL (WWSN)
       (Y OR N) : Y
CHANGE INSTRUMENT OR ATTENUATION OPTIONS ? (Y OR N) : N
MODIFY SOURCE ? (Y OR N) : N
A NEW DISTANCE OR AZIMUTH ? (Y OR N) : Y
DISTANCE, AZIMUTH :4000,67.5
 DISTANCE =    4000.00       AZIMUTH =      67.50
INCLUDE EXTERNAL EXPLOSION ? (Y OR N) : N
ONE SIDED OR DEGRADED SOURCE POTENTIALS ? (Y OR N) : N
ATTENUATION ? (Y OR N) : N
CHANGE ATTENUATION ? (Y OR N) : N
INSTRUMENT ? (Y OR N) : Y
HIGHER MODE ? (Y OR N) : N
INITIAL OR RESETTING INITIAL SEISMOGRAM PARAMETERS ? (Y OR N) : N
LFFT =      3 NT =     512
INITIAL OR RESETTING INITIAL SEISMOGRAM SCALES ? (Y OR N) : N
```

```
        POINT DOUBLE COUPLE OR DISLOCATION
        SOURCE DEPTH =       0.300000
ALPS =  6.2000 ,BETS =  3.5110 ,RHOS =  2.7360 ,MUS = 33.7271
        DELTA = 45.00 ,LAMBDA = 90.00
        DISTANCE =      4000.00      AZIMUTH =       67.5000
BRUNE FAR FIELD BODY WAVE RISE TIME =   0.5200 SECS

  CIT 109  RAYLEIGH FUNDAMENTAL (WWSN)
```

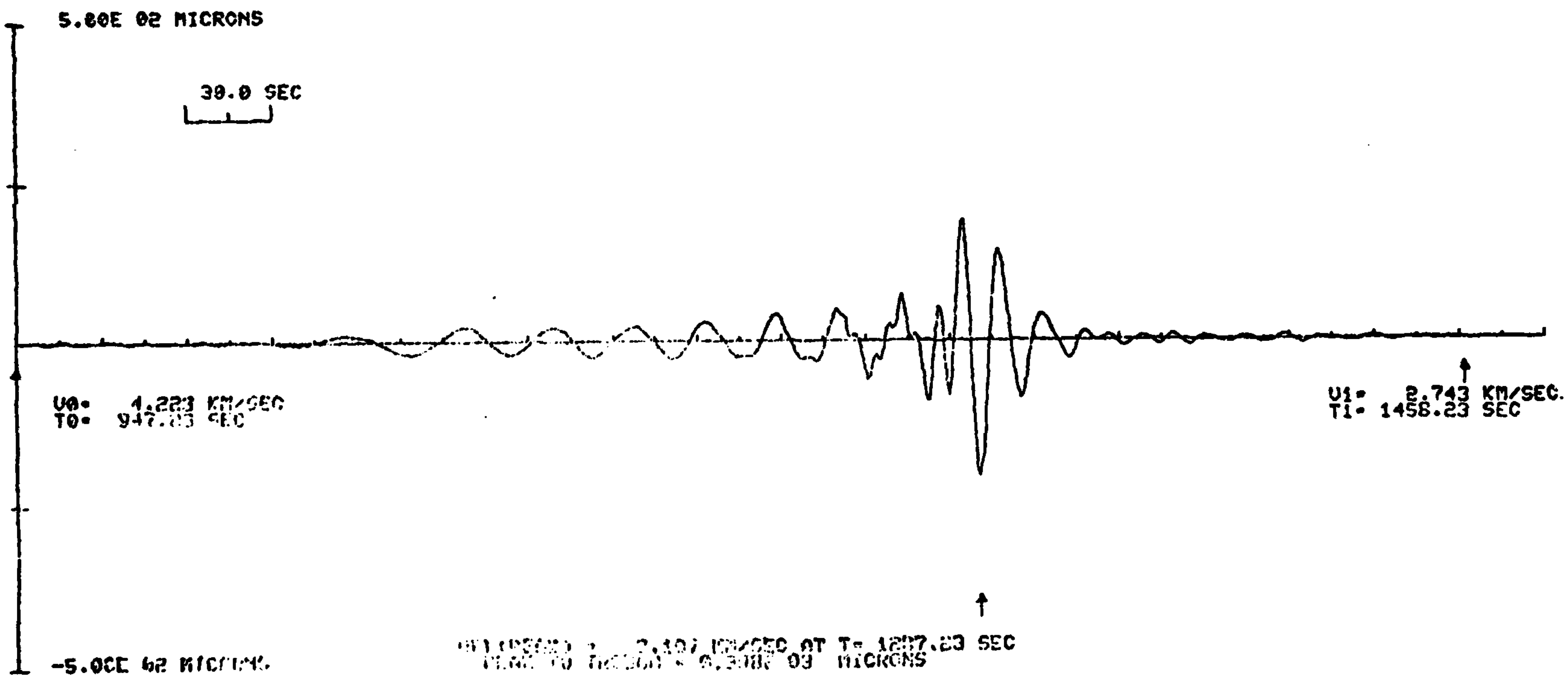
$M_0 = 10^{25}$



SAVE SEISMOGRAM ? (Y OR N) :

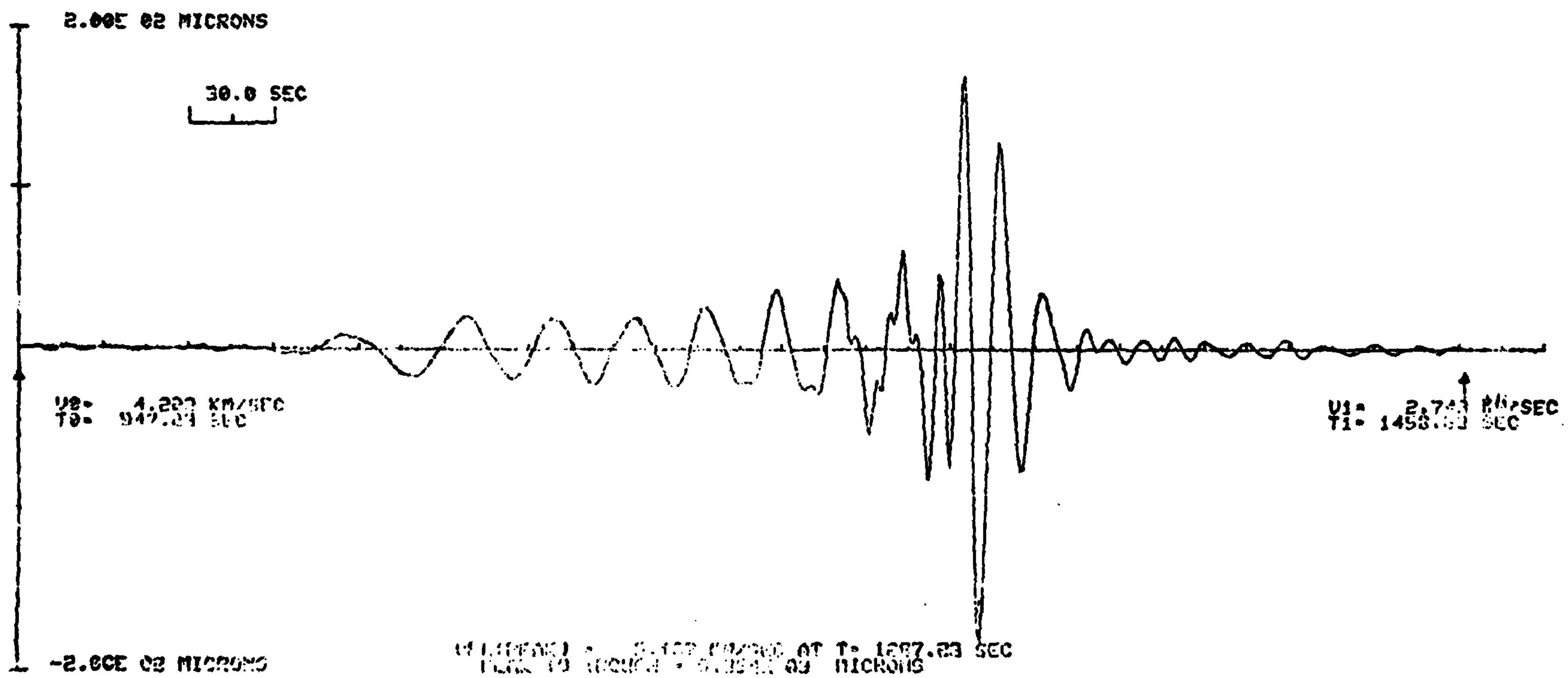
```
RESET SEISMOGRAM SCALES ? (Y OR N) : N
CHANGE SEISMOGRAM PARAMETERS ? (Y OR N) : N
CHANGES TO THIS SEISMOGRAM ?
  CIT 109  RAYLEIGH FUNDAMENTAL (WWSN)
       (Y OR N) : Y
CHANGE INSTRUMENT OR ATTENUATION OPTIONS ? (Y OR N) : N
MODIFY SOURCE ? (Y OR N) : N
A NEW DISTANCE OR AZIMUTH ? (Y OR N) : Y
DISTANCE, AZIMUTH :4000,90
 DISTANCE =   4000.00      AZIMUTH =     90.00
INCLUDE EXTERNAL EXPLOSION ? (Y OR N) : N
ONE SIDED OR DEGRADED SOURCE POTENTIALS ? (Y OR N) : N
ATTENUATION ? (Y OR N) : Y
CHANGE ATTENUATION ? (Y OR N) : N
INSTRUMENT ? (Y OR N) : Y
HIGHER MODE ? (Y OR N) : N
INITIAL OR RESETTING INITIAL SEISMOGRAM PARAMETERS ? (Y OR N) : N
LFFT =      3 NT =     512
INITIAL OR RESETTING INITIAL SEISMOGRAM SCALES ? (Y OR N) : N
```

```
         POINT DOUBLE COUPLE OR DISLOCATION
         SOURCE DEPTH =     0.300000
ALPS =  6.2000 ,BETS =  3.5110 ,RHOS =  2.7360 ,MUS = 33.727
         DELTA = 45.00 ,LAMBDA = 90.00
         DISTANCE =      4000.00      AZIMUTH =      90.0000
BRUNE FAR FIELD BODY WAVE RISE TIME =   0.5000 SECS

 CIT 109  RAYLEIGH FUNDAMENTAL (WWSN)
```



```
RESET SEISMOGRAM SCALES ? (Y OR N) : N                                  UIT,
CHANGE SEISMOGRAM PARAMETERS ? (Y OR N) : N
CHANGES TO THIS SEISMOGRAM ?
  CIT 109  RAYLEIGH FUNDAMENTAL (WWSN)
         (Y OR N) : Y
CHANGE INSTRUMENT OR ATTENUATION OPTIONS ? (Y OR N) : N
MODIFY SOURCE ? (Y OR N) : N
A NEW DISTANCE OR AZIMUTH ? (Y OR N) : Y
DISTANCE, AZIMUTH :4000,112.5
 DISTANCE =   4000.00        AZIMUTH =    112.50
INCLUDE EXTERNAL EXPLOSION ? (Y OR N) : N
ONE SIDED OR DEGRADED SOURCE POTENTIALS ? (Y OR N) : N
ATTENUATION ? (Y OR N) : Y
CHANGE ATTENUATION ? (Y OR N) : N
INSTRUMENT ? (Y OR N) : Y
HIGHER MODE ? (Y OR N) : N
INITIAL OR RESETTING INITIAL SEISMOGRAM PARAMETERS ? (Y OR N) : N
LFFT =     3 NT =    512
INITIAL OR RESETTING INITIAL SEISMOGRAM SCALES ? (Y OR N) : N
```

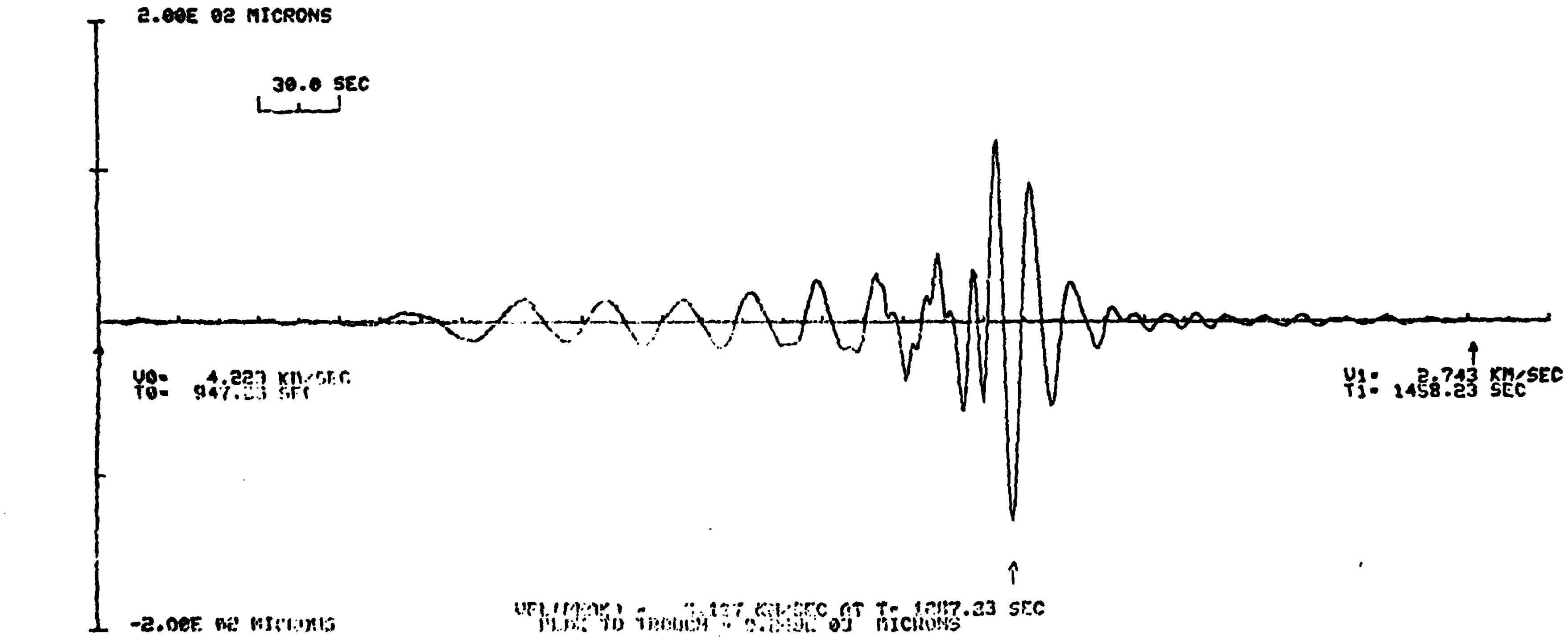
```
      POINT DOUBLE COUPLE OR DISLOCATION
      SOURCE DEPTH =      0.300000
ALPS =  6.2000 ,BETS =  3.5110 ,RHOS =  2.7360 ,MUS = 33.7270
      DELTA = 45.00 ,LAMBDA = 90.00
      DISTANCE =       4000.00     AZIMUTH =      112.500
BRUNE FAR FIELD BODY WAVE RISE TIME =   0.5000 SECS

 CIT 109  RAYLEIGH FUNDAMENTAL (WWSN)
```



SAVE SEISMOGRAM ? (Y OR N) :

POINT DOUBLE COUPLE OR DISLOCATION
SOURCE DEPTH = 0.300000
ALPS = 6.2000 ,BETS = 3.5110 ,RHOS = 2.7360 ,MUS = 33.7270
DELTA = 45.00 ,LAMBDA = 90.00
DISTANCE = 4000.00 AZIMUTH = 135.000
BRUNE FAR FIELD BODY WAVE RISE TIME = 0.5000 SECS

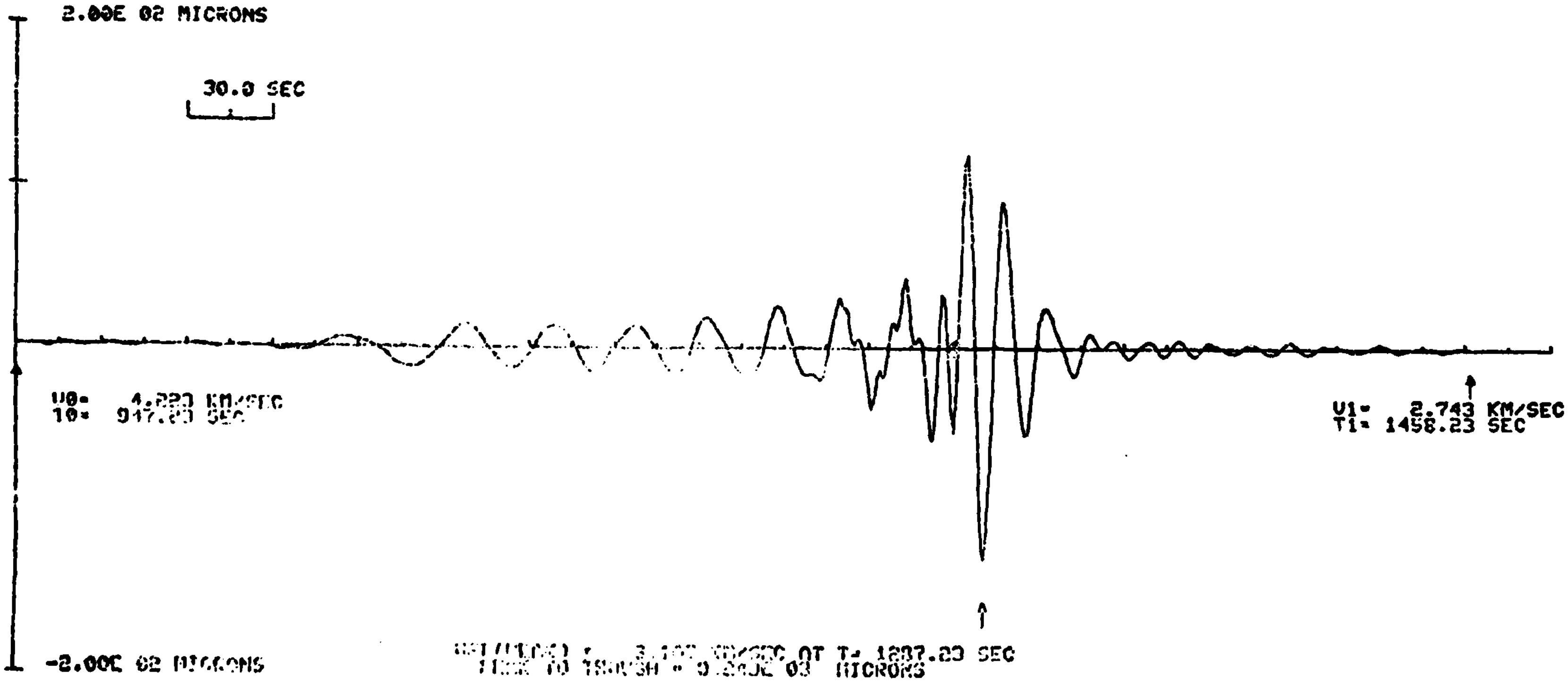
CIT 109 RAYLEIGH FUNDAMENTAL (WWSN)



SAVE SEISMOGRAM ? (Y OR N) :

```
RESET SEISMOGRAM SCALES ? (Y OR N) : N
CHANGE SEISMOGRAM PARAMETERS ? (Y OR N) : N
CHANGES TO THIS SEISMOGRAM ?
  CIT 109  RAYLEIGH FUNDAMENTAL (WWSN)
         (Y OR N) : Y
CHANGE INSTRUMENT OR ATTENUATION OPTIONS ? (Y OR N) : N
MODIFY SOURCE ? (Y OR N) : N
A NEW DISTANCE OR AZIMUTH ? (Y OR N) : Y
DISTANCE, AZIMUTH :4000,225
 DISTANCE =    4000.00        AZIMUTH =     225.00
INCLUDE EXTERNAL EXPLOSION ? (Y OR N) : N
ONE SIDED OR DEGRADED SOURCE POTENTIALS ? (Y OR N) : N
ATTENUATION ? (Y OR N) : Y
CHANGE ATTENUATION ? (Y OR N) : N
INSTRUMENT ? (Y OR N) : Y
HIGHER MODE ? (Y OR N) : N
INITIAL OR RESETTING INITIAL SEISMOGRAM PARAMETERS ? (Y OR N) : N
LFFT =      3 NT =     512
INITIAL OR RESETTING INITIAL SEISMOGRAM SCALES ? (Y OR N) : N
```

```
      POINT DOUBLE COUPLE OR DISLOCATION
      SOURCE DEPTH =        0.300000
ALPS =  6.2000 ,BETS =  3.5110 ,RHOS =  2.7360 ,MUS = 33.7270
      DELTA = 45.00 ,LAMBDA = 90.00
      DISTANCE =         4000.00      AZIMUTH =       225.000
BRUNE FAR FIELD BODY WAVE RISE TIME =   0.5000 SECS

CIT 109  RAYLEIGH FUNDAMENTAL (WWSN)
```

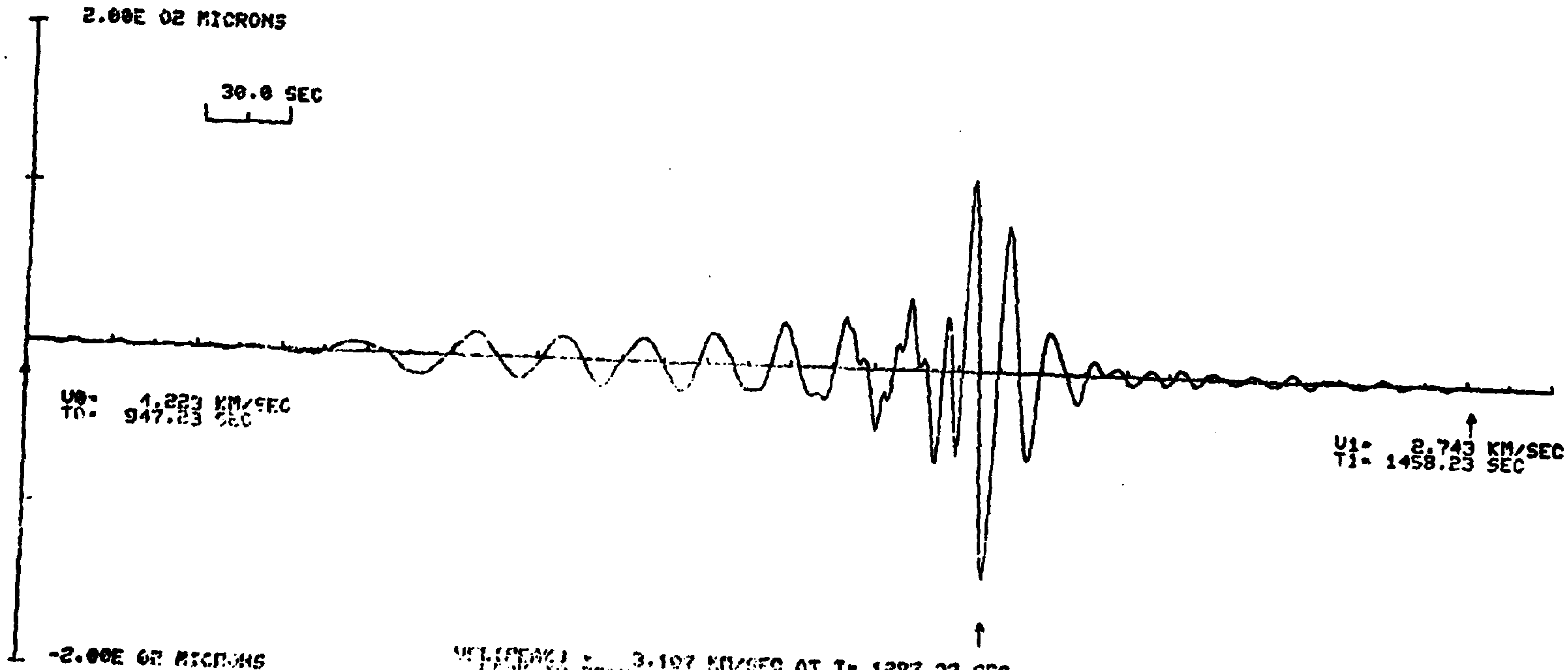



```
SAVE SEISMOGRAM ? (Y OR N) :
```

```
RESET SEISMOGRAM SCALES ? (Y OR N) : N
CHANGE SEISMOGRAM PARAMETERS ? (Y OR N) : N
CHANGES TO THIS SEISMOGRAM ?
  CIT 109  RAYLEIGH FUNDAMENTAL (WWSN)
        (Y OR N) : Y
CHANGE INSTRUMENT OR ATTENUATION OPTIONS ? (Y OR N) : N
MODIFY SOURCE ? (Y OR N) : N
A NEW DISTANCE OR AZIMUTH ? (Y OR N) : Y
DISTANCE, AZIMUTH :4000.,315
 DISTANCE =    4000.00        AZIMUTH =      315.00
INCLUDE EXTERNAL EXPLOSION ? (Y OR N) : N
ONE SIDED OR DEGRADED SOURCE POTENTIALS ? (Y OR N) : N
ATTENUATION ? (Y OR N) : Y
CHANGE ATTENUATION ? (Y OR N) : N
INSTRUMENT ? (Y OR N) : Y
HIGHER MODE ? (Y OR N) : N
INITIAL OR RESETTING INITIAL SEISMOGRAM PARAMETERS ? (Y OR N) : N
LFFT =      3 NT =      512
INITIAL OR RESETTING INITIAL SEISMOGRAM SCALES ? (Y OR N) : N
```

PAUSE

```
            POINT DOUBLE COUPLE OR DISLOCATION
            SOURCE DEPTH =      0.300000
ALPS =   6.2000 ,BETS =   3.5110 ,RHOS =   2.7360 ,MUS = 33.7270
            DELTA = 45.00 ,LAMBDA = 90.00
            DISTANCE =        4000.00        AZIMUTH =       315.000
BRUNE FAR FIELD BODY WAVE RISE TIME =    0.5000 SECS

CIT 109  RAYLEIGH FUNDAMENTAL (WWSN)
```



SAVE SEISMOGRAM ? (Y OR N) : N

AFTAC

TECTONIC EFFECTS

Thomas D. Eisenhauer

The cause of so-called "tectonic effects" is still a matter of speculation. The effects are of concern because they modify and sometimes dominate the explosion generated Rayleigh waves. When tectonic effects result in signicant modification of LR, any yield estimates from LR are unreliable. At present, no generally accepted method is available to remove tectonic effects from the composite Rayleigh waves accompanying an explosion. It is important to identify explosions for which tectonic effects are pronounced enough to cause problems with the explosion generated Rayleigh waves.

The best evidence for tectonic interference comes from analysis of the Rayleigh wave amplitudes and phase. Asymmetric radiation and/or reversals in phase are evidence for interference. In general, path properties are not known well enough to determine asymmetric radiation or initial phase for LR from a single explosion observed at teleseismic distances. When explosions at a test site exhibit various levels of interference it is possible to detect that something is wrong by analyzing LR, but it is still difficult to identify which events have the strongest interference. Most hypotheses used to explain "tectonic effects" also explain the excitation of shear motion. For explosions of a given yield and with identical orientation of secondary movements at the source, the larger the shear wave amplitude, the greater the potential for Rayleigh wave interference. With this oversimplification, shear wave amplitude can be used to determine the relative levels of tectonic effects for explosions. The same line of reasoning leads to the use of Love wave amplitudes to identify the potential for tectonic interference.

When the ratio of horizontal to vertical movements at the source changes, the shear and Love wave amplitudes are not valid indicators of relative levels of interference in LR. We have very little information on the nature of explosion generated tectonic effects in different tectonic settings. The strike-slip representation invoked for the Nevada test site represents one of many possible types of orientation of movements. If thrust faulting can produce a reversal of LR phase in all quadrants and reduce explosion generated LR, then normal faulting should result in the opposite effect. I believe that much more research is required to develop an improved M_S - yield estimation procedure which compensates for tectonic effects.

References

Marshall, P. D., D. L. Springer, and H. C. Rodean, "Magnitude Corrections for Attenuation in the Upper Mantle", 1977, Preprint UCRL-80423.

Murphy, J. R. 1977, "Seismic Coupling and Magnitude/Yield Relations for Underground Nuclear Detonations in Salt, Granite, Tuff/Rhyolite and Shala Emplacement Media," Computer Sciences Corporation TR CSC TR-77-0004.

Veith, K. F., and G. E. Clawson, 1972, "Magnitude from Short-Period P-Wave Data," Bull. Seism. Soc. Am., 6, 435-452.

EFFECTS OF ATTENUATION ON SURFACE WAVES

Keith Priestley
Institute of Geophysics and Planetary Physics
Scripps Institution of Oceanography
University of California, San Diego
La Jolla, California 92093

Surface wave group and phase velocities have been studied extensively to make inferences concerning the elastic properties of the earth as a function of depth. The decay of surface wave amplitudes with distance are less frequently studied due to the difficulty in making meaningful determinations. Difficulties arise in determining surface wave attenuation because the observed amplitudes are affected by scattering, multipathing, lateral refraction, mode conversion, source parameters, local geology, instrumental effects in addition to intrinsic anelasticity. Hence, surface wave attenuation data are less abundant and subject to greater observational uncertainties than surface wave velocity data.

Presently available surface wave attenuation data obtained for paths entirely within either continent or ocean regions include eastern North America (Mitchell, 1973a, 1973b; Nuttli, 1973; Herrman and Mitchell, 1975; Mitchell, 1975, Street, 1976; and Bollinger, 1979), Western North America (Mitchell, 1975; Lee and Solomon, 1975, 1978, 1979), the Pacific Basin (Mitchell, 1976; Mitchell *et al.*, 1976) and Eurasia (Chen and Molnar, 1975; Yacoub and Mitchell, 1977). The majority of these results have been summarized by Mitchell *et al.* (1977). Attenuation data for eastern North America are more numerous than for other areas of the world.

Hermann and Mitchell (1975) determined attenuation coefficients for Central and Eastern North America from surface waves of eight well recorded events. Representative values are given in Table 1. Mitchell (1975) analyzed seismograms of surface waves generated by two underground nuclear explosions in Colorado and found significant regional differences in attenuation between Eastern and Western North America at short periods (<16 seconds). They found little difference in attenuation

TABLE 1. Amplitude Ratios and Attenuation Coefficient Differences for Rayleigh Waves Recorded in Eastern and Western North America.

Period, s	Amplitude Ratio West/East	$\gamma_W - \gamma_E$, 10^{-4} km^{-1}	γ_E, 10^{-4} km^{-1}	γ_W, 10^{-4}, km^{-1}
5	0.19	16.60	5.10	21.70
6	0.22	15.10	5.50	20.60
7	0.31	11.70	5.87	17.60
8	0.53	6.35	4.06	10.40
9	0.60	5.11	2.45	7.56
10	0.77	2.61	3.12	5.73
12	0.76	2.74	2.70	5.44
14	0.79	2.36	1.37	3.73
16	0.85	1.63	1.76	3.39
18	0.86	1.51	1.01	2.52
20	0.87	1.39	1.19	2.58
25	1.00	0.00	1.28	1.28
30	0.97	0.30	1.31	1.61
35	0.96	0.41	0.75	1.16
40	0.91	0.94	1.25	2.19

between Eastern and Western North America at longer periods (20 to 35 seconds). These two period ranges correspond to energy propagating in the upper crust and the whole crust, respectively indicating the major difference in attenuation occurs in the upper crust. Lee and Solomon

(1979) find a zone of high seismic attenuation in the upper mantle beneath both Eastern and Western North America. However, this zone is more highly attenuating and occurs at shallower depths in Western North America than in Eastern North America. Determination of surface wave attenuation for paths across Eurasia are currently more difficult than North America because data acquisition is more difficult. Yacoub and Mitchell (1977) present Rayleigh wave attenuation data for Eurasia using six earthquakes and two nuclear explosions. The mean Rayleigh wave attenuation coefficient values found for the stable regions of Eurasia are nearly the same as those obtained for Eastern North America while the attenuation coefficients determined for the tectonically active portions of Eurasia are similar to those found for Western North America. This result implies that the distribution of Q_B^{-1} with depth in the crust of both regions cannot be greatly different. Mitchell *et al.* (1976) compiled surface wave attenuation data for the Pacific Ocean. These values at periods greater than 20 seconds are not significantly different from those determined for continental areas.

The variation of surface wave attenuation with frequency primarily results from the fact that the various periods sample different parts of the earth's structure, but also include any effects of frequency dependent Q. At periods of 20 seconds and greater, the variation of attenuation coefficient with period is similar (to within 30%) in all regions, hence, the success of the Gutenberg (1945) empirical relationship for surface wave magnitude based on the amplitude of 20 second surface waves. At periods less than 20 seconds, surface waves are composed of energy propagating primarily in the upper crust where the structure and

attenuation are much more regionally dependent, hence requiring a much more extensive evaluation of the transmission properties for individual paths.

REFERENCES

Bollinger, G. A., Attenuation of the Lg phase and the determination of m_b in the southeastern United States, Bull. Seism. Soc. Amer., 69, 45-63, 1979.

Chen, W. P., Molnar, P., Short-period Rayleigh-wave dispersion across the Tibetan Plateau, Bull. Seism. Soc. Amer., 65, 1051-1058, 1975.

Gutenberg, G., Amplitude of surface wave magnitudes of shallow earthquakes, Bull. Seism. Soc. Amer., 35, 3-12, 1945.

Hermann, R. B., and B. J. Mitchell, Statistical analysis and interpretation of surface wave anelastic attenuation data for the stable interior of North America, Bull. Seism. Soc. Amer., 65, 1115-1128, 1975.

Lee, W. B., and S. C. Solomon, Inversion schemes for surface wave attenuation and Q in the crust and the mantle, Geophys. J. Roy. Astron. Soc., 43, 47-71, 1975.

Lee., W. B., Solomon, S. C., Simultaneous inversion of surface wave phase velocity and attenuation: Love waves in Western North America, J. Geophys. Res., 83, 3389-3400, 1978.

Lee., W. B., Solomon, S. C. Simulatneous inversion of surface-wave velocity and attenuation: Rayleigh and Love waves over continental and oceanic paths, Bull. Seism. Soc. Amer., 69, 65-96, 1979.

Mitchell, B. J., Radiation and attenuation of Rayleigh waves from the southeastern Missouri earthquake of October 21, 1965, J. Geophys. Res., 78, 886-899, 1973a.

Mitchell, B. J., Surface wave attenuation and crustal anelasticity in central North America, Bull. Seism. Soc. Amer., 63, 1057-1071, 1973b.

Mitchell, B. J., Regional Rayleigh wave attenuation in North America, J. Geophys. Res., 80, 4904-4916, 1975.

Mitchell, B. J., Anelasticity of the crust and upper mantle beneath the Pacific Ocean from the inversion of observed surface wave attenuation. Geophys. J. Roy. Astron. Soc., 46, 521-533, 1976.

Mitchell, B. J., L. W. B. Leite, Y. K. Yu, and R. B. Herrmann, Attenuation of Love and Rayleigh waves across the Pacific at periods between 15 and 110 seconds, Bull. Seism. Soc. Amer., 66, 1189-1201, 1976.

Mitchell, B. J. *et al.*, The Earth's Crust, Mono. 20, Amer. Geophys. Union, Washington D. C., 1977.

Nuttli, O. W., Seismic wave attenuation and magnitude relations for eastern North America, J. Geophys. Res., 78, 876-885, 1973.

Street, R. L., Scaling northeastern United States/southeastern Canadian earthquakes by their Lg waves, Bull. Seims. Soc. Amer., 66, 1525-1537, 1976.

Yacoub, N. K., and B. J. Mitchell, Attenuation of Rayleigh wave amplitudes across Eurasia, Bull. Seism. Soc. Amer., 67, 751-769, 1977.

St. Louis University

EFFECTS OF ATTENUATION ON SURFACE WAVES

Otto W. Nuttli

Surface-wave attenuation depends on three factors, namely geometric spreading, dispersion and absorption. The first two factors are controlled by the shear-wave velocity distribution in the crust and mantle, and the third by the Q structure. Furthermore, the first two factors are dominant at the shorter epicentral distances, and the third at the larger distances, because the first two are of the form $(\sin \Delta)^{-n}$ and Δ^{-n}, and the third is of the form $\exp(-\gamma\Delta)$.

For 20-sec period Rayleigh waves the value of γ is 0.015 deg^{-1} (0.000135 km^{-1}) for a world-wide average. This leads to the standard M_S formula

$$M_S = 3.30 + 1.66 \log \Delta \text{ (deg)} + \log A/T \text{ (microns/sec)} \qquad (1)$$

which is a linear approximation to the Rayleigh-wave attenuation curve for distances of 25° to 140° for 20-sec period waves. Nuttli and Kim (1975) showed that if the distance interval is 10° to 25°, the appropriate formula for 20-sec period Rayleigh waves is

$$M_S = 4.16 + 1.07 \log \Delta \text{ (deg)} + \log A/T \text{ (microns/sec)}. \qquad (2)$$

Formula (2) has advantages over regional formulas which use shorter-period surface waves to estimate M_S, such as those of Karnik et al (1962), Basham (1971), Evernden et al (1971), Marshall and Basham (1972) and Nuttli (1973). First, the absorption of 20-sec period surface waves depends on the Q of the bottom of the lithosphere, which exhibits no large regional variation. The shorter-period surface waves used for the regional formulas, on the other hand, depend on the Q structure of the upper crust, which is

highly variable. Second, M_S is defined in terms of amplitudes of 20-sec period surface waves, so there is no problem in spectral scaling when using formula (2), which also utilizes 20-sec period waves. The regional phase formulas, however, because they involve waves of period 3 to 15 sec, which do not scale in the same way as 20-sec period waves for all magnitudes, are limited in their applicability to a particular range of M_S values.

REFERENCES

Basham, P.W. (1971) A new magnitude formula for short period continental Rayleigh waves, Geophys. J., 23, 255-266.

Evernden, J.F., W.J. Best, P.W. Pomeroy, T.V.McEvilly, J.M. Savino and L.R. Sykes (1971) Discrimination between small-magnitude earthquakes and explosions, J. Geophys. Res., 76, 8042-8055.

Karnik, V., N.V. Kondorskaya, Y.V. Riznichenko, E.F. Savarensky, S.L. Soloviev, N. V. Shevalin, J. Vanek and A. Zatopek (1962) Standardization of the earthquake magnitude scale, Studia Geophys. Geodaet., 6, 41-47.

Marshall, P.D. and P.W. Basham (1972) Discrimination between earthquake and underground explosions employing an improved M_S scale, Geophys. J., 28, 431-458.

Nuttli, O.W. (1973) Seismic wave attenuation and magnitude relations for eastern North America, J. Geophys. Res., 78, 876-885.

Nuttli, O.W. and S.G. Kim (1975) Surface-wave magnitudes of Eurasian earthquakes and explosions, Bull. Seism. Soc. Am., 65, 693-709.

Effects of Attenuation
on Surface Waves

by

Brian J. Mitchell
Department of Earth and Atmospheric Sciences
Saint Louis University

Introduction

This report will summarize recent results on seismic surface wave attenuation at intermediate periods (4-50s in continents and 15-110s oceanic regions) and will present models of shear wave internal friction (Q_{β}^{-1}) which satisfy those data in several regions of the world. In addition to summarizing research on regional variations of Rayleigh wave attenuation, recent results on the frequency dependence of Q in the continental crust will be discussed. Finally, the regionalized Q_{β}^{-1} models of the crust and upper mantle obtained from surface waves will be used to estimate values of t* and Q for shear waves traversing them.

Some of the work in the following sections is still in progress, or manuscripts have only recently been submitted for publication. Figures are included in this report only to present the results of this more recent work which has not yet appeared in the open literature. The results of older published studies will be referred to in the text.

In the course of several of the studies discussed below, new methods have been developed to determine surface wave attenuation coefficient values and models of Q_{β}^{-1} for the crust and upper mantle. The present report will, however, concentrate on results obtained and will discuss methods only briefly.

Regional Variations of Surface Wave Attenuation

Attenuation data have been obtained for average world-circling paths, long paths across laterally varying structures, and relatively short paths over regions which can be considered to be nearly laterally uniform. It is only through studies of the last of these three types which will provide detailed information on the lateral variation of Q structure in the Earth. The following sections will concentrate on results of surface wave attenuation studies over regions which will be considered to be relatively uniform laterally. Results of regional Rayleigh wave attenuation studies have been completed for both continents and oceans. Love wave data have also been obtained for those regions; however the large scatter usually present in Love wave attenuation data makes it more difficult to use those waves to study regional variations in attenuation. The following discussions will therefore be restricted to studies of Rayleigh wave attenuation.

Continental regions. The most extensive available set of continental surface wave attenuation data has been collected for eastern North America (Mitchell, 1973 a, b; Hasegawa, 1973 ; Herrmann and Mitchell, 1975 ; Lee and Solomon, 1975). The agreement of the attenuation coefficient values obtained by various investigators is quite good. Mitchell (1973b) and Herrmann and Mitchell (1975) inverted their attenuation data to obtain models of Q_{β}^{-1} for the crust of eastern North America. Lee and Solomon (1975) inverted longer period data to obtain a Q_{β}^{-1} of the crust and upper mantle for the eastern United States. The data of Herrmann and

Mitchell (1975) yielded a crustal model in which Q_β was about 250 in the upper crust and much larger (1000-2000) in the lower crust.

Mitchell (1975), in a study of Rayleigh waves from the nuclear events Rulison and Rio Blanco in Colorado, concluded that Rayleigh wave attenuation in the crust of western North America is substantially greater than that in eastern North America. Those attenuation data could be explained if Q_β in the upper crust of western North America is about half as large as that in the upper crust of eastern North America. Lee and Solomon (1975, 1978) inverted relatively long-period data to obtain Q_β models of the upper mantle in the western United States. Their western United States model included much lower Q values in the upper mantle than did their eastern United States model.

Attenuation coefficients in the studies cited above were obtained either by (1) the two-station method or (2) comparing observed and theoretically predicted amplitude radiation patterns (Tsai and Aki, 1969). The former method requires two stations along a great circle path from a suitable earthquake epicenter. The latter method requires several stations surrounding, and lying at varying distances from an earthquake with a known depth and fault-plane solution. These methods require relatively long paths or large regions over which to obtain attenuation coefficient values.

In order to study the anelastic properties of smaller tectonic regions such as the Basin and Range province or Colorado Plateau in the western United States, a new method has been developed which employs only one earthquake and one recording station within the region of interest (Cheng and Mitchell, 1978; Cheng, 1980).

The method employs the fundamental mode as well as higher modes of Rayleigh waves from an earthquake of known focal depth and fault-plane solution. A difficulty in using the method arises because the higher modes overlap in group velocity over a considerable period range. Consequently, our observed amplitudes correspond to a superposition of several higher modes rather than a single mode. Since we could not separate the higher modes, we compared our observations with theoretical values computed for a superposition of several modes. An example of the observed and theoretical spectra for a path across the Basin and Range Province appears in Figure 1. The method consists of obtaining the best fit between observed and theoretical amplitude spectra of the fundamental and higher modes, simultaneously. Cheng (1980) has devised a stochastic inversion method which solves for the Q_β model and earthquake seismic moment which explain the observed spectral data.

Figure 2 presents Q_β models for the eastern United States, the Colorado Plateau, and the Basin and Range province which have been obtained using the newly developed method. The eastern United States model is very similar to that obtained by Herrmann and Mitchell (1975) using only fundamental-mode data. The Basin and Range model is characterized by very low Q values in the upper crust, and the Colorado Plateau lies between the other models. The lower crust in all models has very high values of Q; however that value is not well determined.

Rayleigh wave attenuation coefficients have also been obtained for Eurasia (Yacoub and Mitchell, 1977). Their data were characterized by large scatter due to sparse station coverage and the necessity

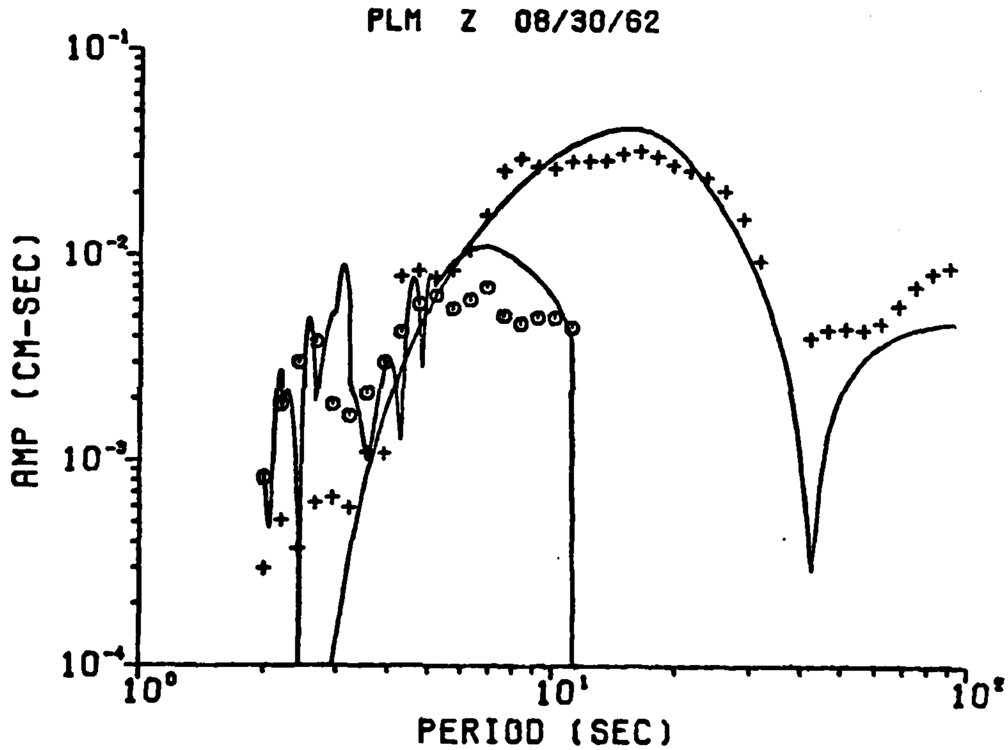


Figure 1. Observed spectra for the fundamental-mode (crosses) and higher-mode (circles) Rayleigh waves across the Basin and Range Province. The theoretical spectra (solid lines) were computed for the Basin and Range model of Figure 2. The higher-mode spectra consist of a superposition of several higher modes. From Cheng (1980).

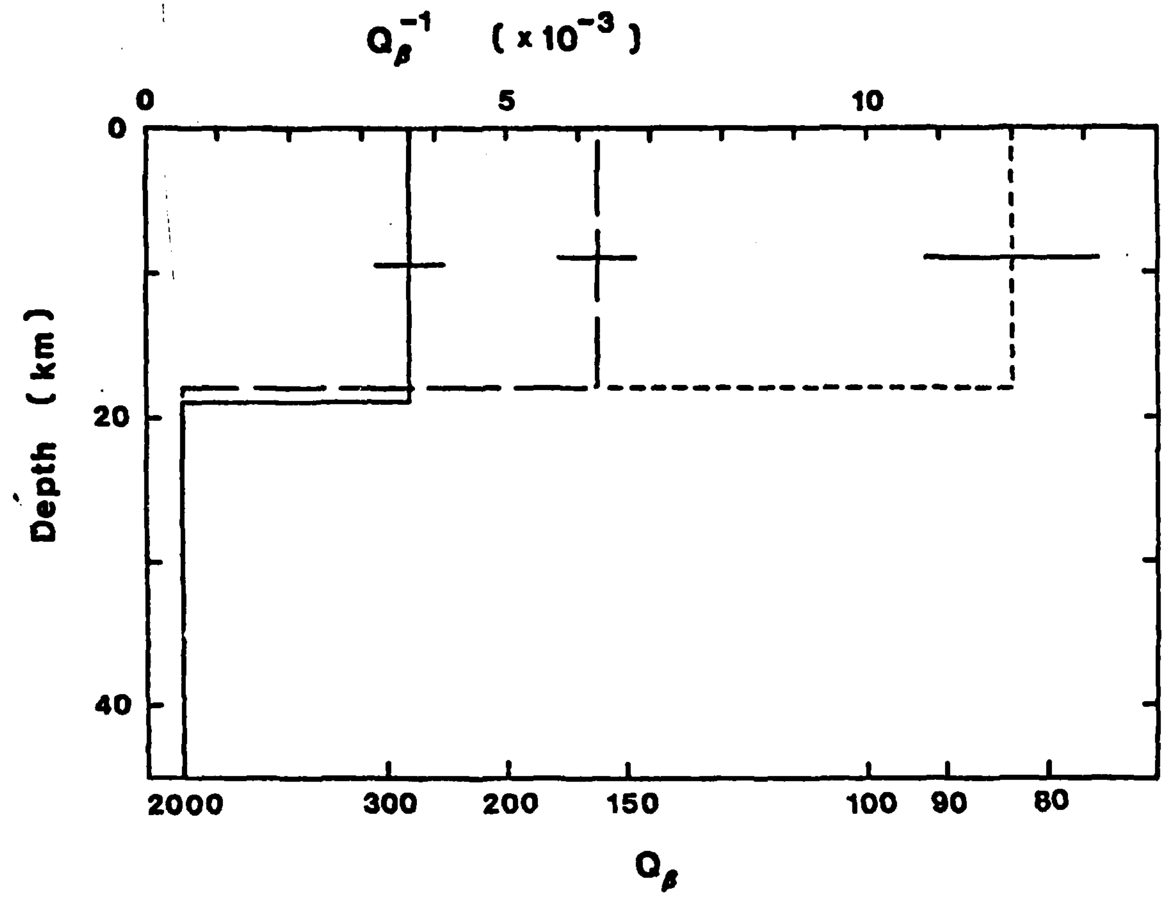


Figure 2. Two-layer Q_β models for the eastern United States (solid line), the Colorado Plateau (long dashes), and the Basin and Range (short dashes), resulting the inversion of fundamental- and higher-mode amplitude spectra. From Cheng (1980).

to use long paths over complex regions. Their results were improved by dividing Eurasia into two large regions which they termed stable and tectonic. Their attenuation coefficeint values for the stable regions are similar to values in eastern North America and their values for the tectonic regions are substantially higher.

Oceanic regions. Mitchell *et al.* (1976) obtained a set of average Rayleigh and Love wave attenuation coefficient data for the Pacific Ocean and Mitchall (1976) inverted those data to obtain a Q^{-1}_{β} model for the crust and upper mantle beneath the Pacific.

Canas and Mitchell (1978) attempted to obtain a data set of high enough quality to detect lateral variations of Q_{β} in the crust and upper mantle across a broad region of the Pacific. They used only island stations to minimize the effect of continental margins and divided to Pacific into three regions according to age (0-50 m.y., 50-100 m.y., and >100 m.y.). They detected a systematic variation of attenuation and Q_{β} with age of the Pacific sea floor, older regions having higher Q values than younger regions.

Mitchell *et al.* (1977) and Correig and Mitchell (1980) noted high values of Rayleigh wave attenuation coefficients in the eastern Pacific, especially along the East Pacific Rise. Q^{-1}_{β} models for these regions have been derived by Canas *et al.* (1980). A summary of Rayleigh wave attenuation coefficients for the Pacific appears in Figure 3 and models for several regions of the Pacific are shown in Figure 4. These results indicate that Q_{β} is very low near the East Pacific Rise, that Q_{β} increases fairly rapidly with distance going

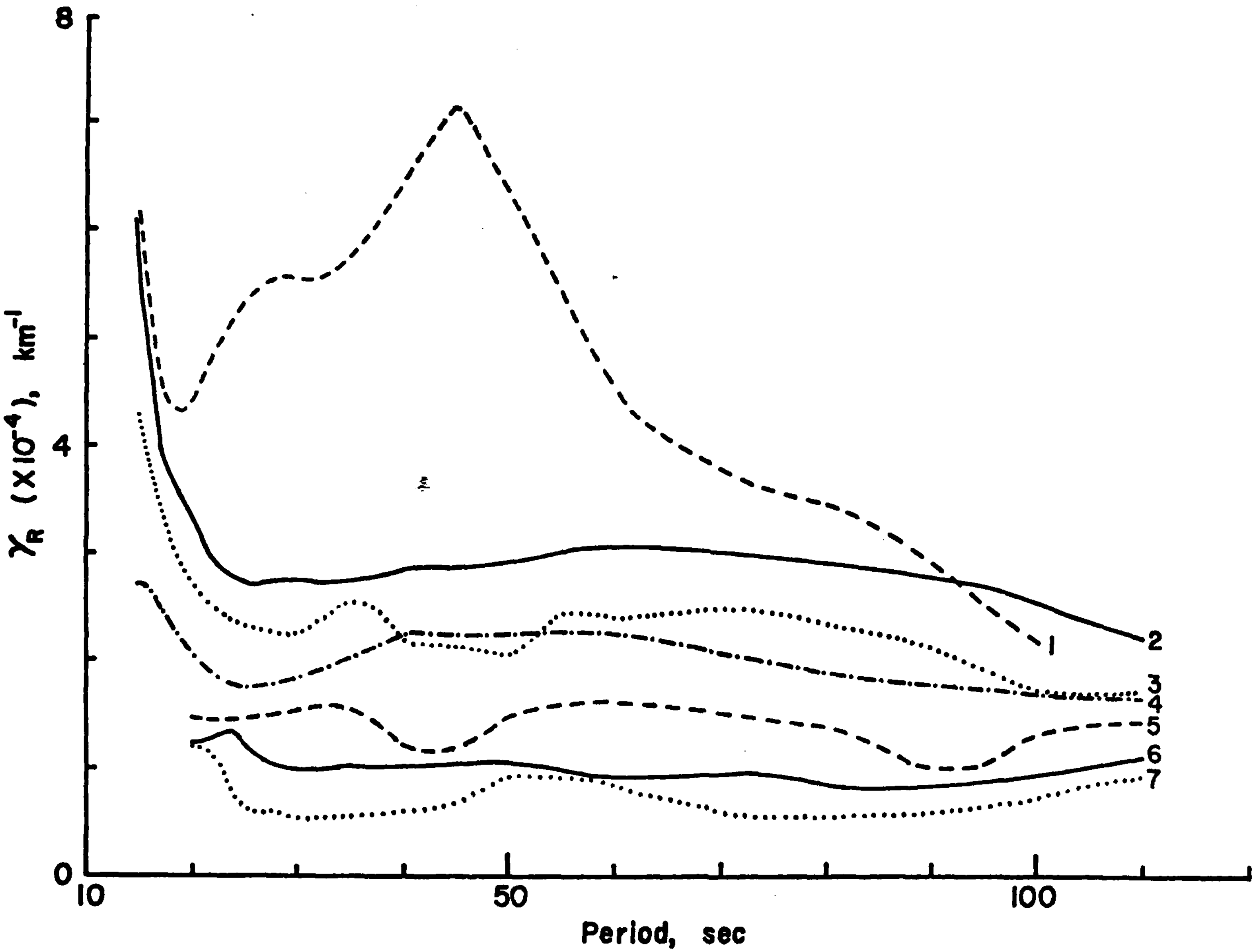


Figure 3. Observed Rayleigh wave attenuation coefficients for (1) the East Pacific Rise, (2) the Nazca plate, (3) the combined Cocos and Nazca plates, (4) the northeastern Pacific (Correig and Mitchell, 1980), and values for regions of the Pacific plate which are (5) 0-50 m.y., (6) 50-100 m.y., and (7) > 100 m.y. in age (Canas and Mitchell, 1978)

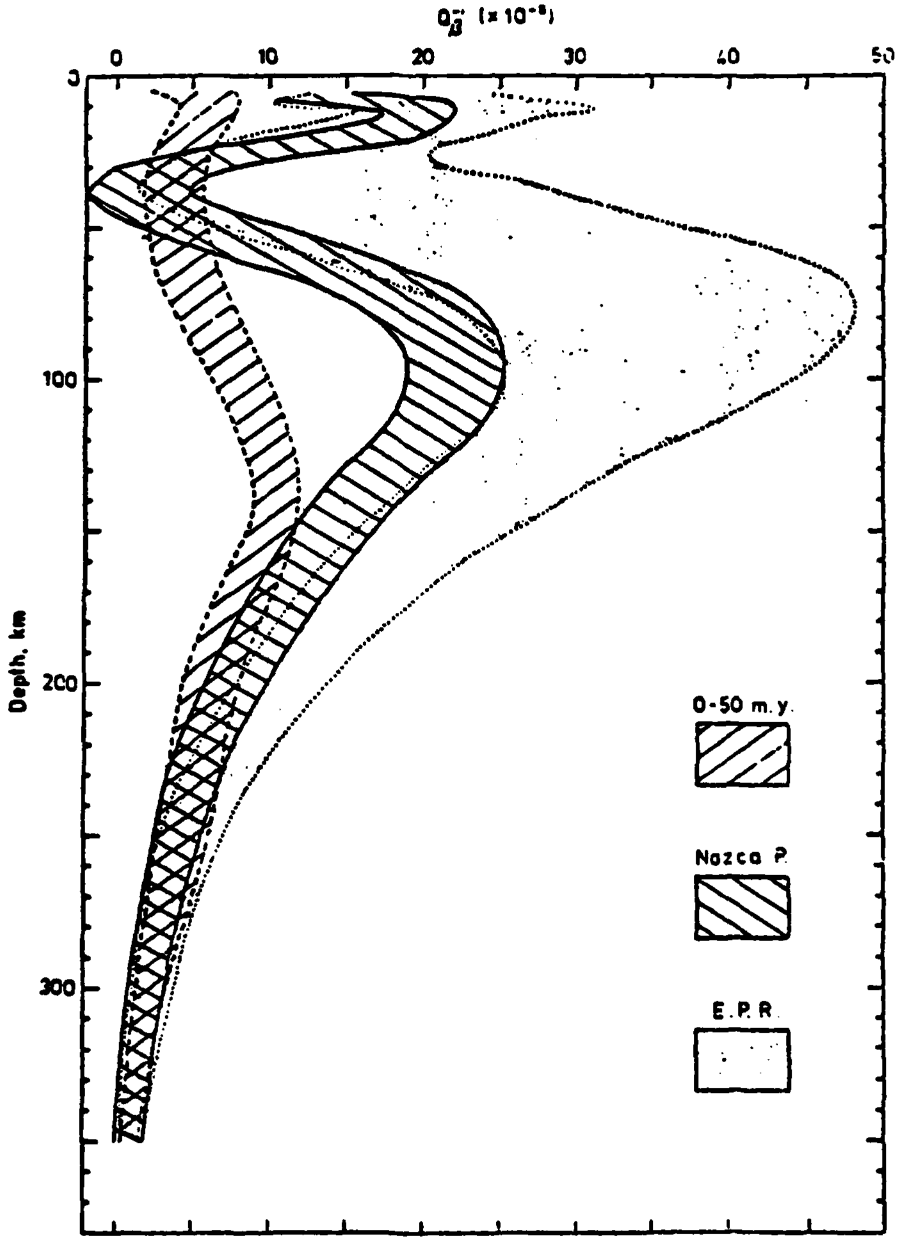


Figure 4. Q_β^{-1} models of the crust and upper mantle beneath the Pacific Ocean. From Canas _et al._ (1980).

away from the ridge, and increases more slowly with distance from the ridge in older regions of the Pacific. Recent work in the Atlantic (Canas, 1980) indicates that Q values beneath the Atlantic also increase with age of the ocean floor (Figure 5). Q values near the ridge, however, are higher than values at corresponding distances from the East Pacific Rise.

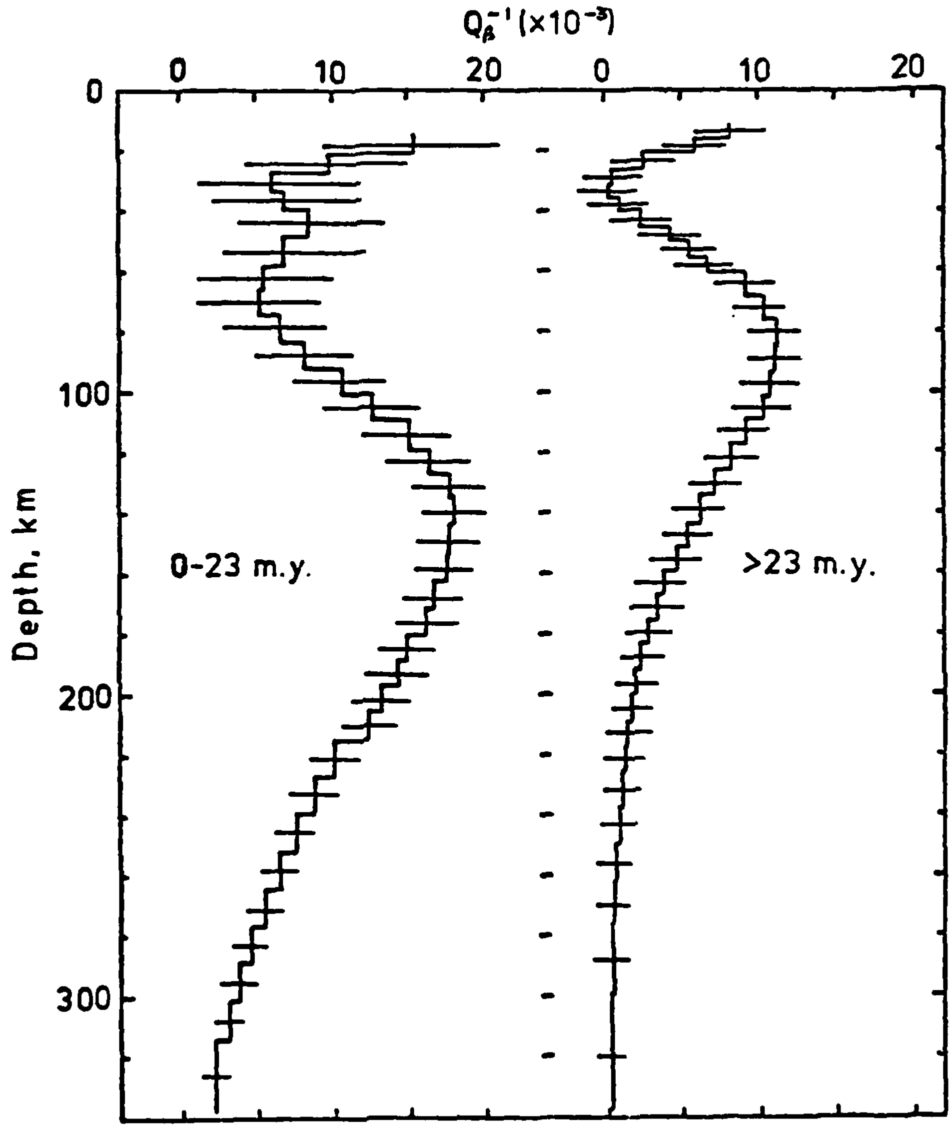


Figure 5. Q_β^{-1} models for the crust and upper mantle beneath two regions of the Atlantic Ocean. From Canas (1980).

Frequency Dependence of Q_β

The inversions in the preceding section assumed that Q_β is independent of frequency. That assumption is made largely because it is thought that the magnitude of frequency dependence in the seismic frequency band is small. A recent study (Mitchell, 1980) used higher- and fundamental-mode data to investigate the nature of Q_β frequency dependence in the crust. Using the fundamental mode alone cannot yield information on the frequency dependence of Q_β since any effect on Rayleigh wave amplitudes due to variation of Q_β with frequency could also be explained by a depth variation of Q_β.

Assuming that $Q^{-1}_\beta(\omega,z) = C(z)\omega^{-\zeta}$, fundamental-mode Rayleigh wave data were inverted to obtain several frequency-dependent Q^{-1}_β models corresponding to different distributions of ζ. Limiting values of ζ were then sought by computing attenuation coefficients for the higher Rayleigh modes for the various models. The higher-mode attenuation data consisted of reported values of Lg (Nuttli, 1973; Street, 1976;) Pomeroy, 1977; Bollinger, 1979) at 1s and values for the first higher Rayleigh mode at periods between 4 and 10s (Mitchell, 1973; Herrmann, 1973) in eastern North America (Figure 6). A comparison of the observed and calculated values suggests that ζ lies betwen 0.3 and 0.5 if it is assumed to be constant over the entire period range between 1 and 40s. It is clear that the frequency-independent case ($\zeta = 0.0$) can be rejected.

Better fits to the data can be achieved by allowing ζ to vary with period, increasing with decreasing period at periods of 4s and less. Several distributions of ζ are possible, partly because

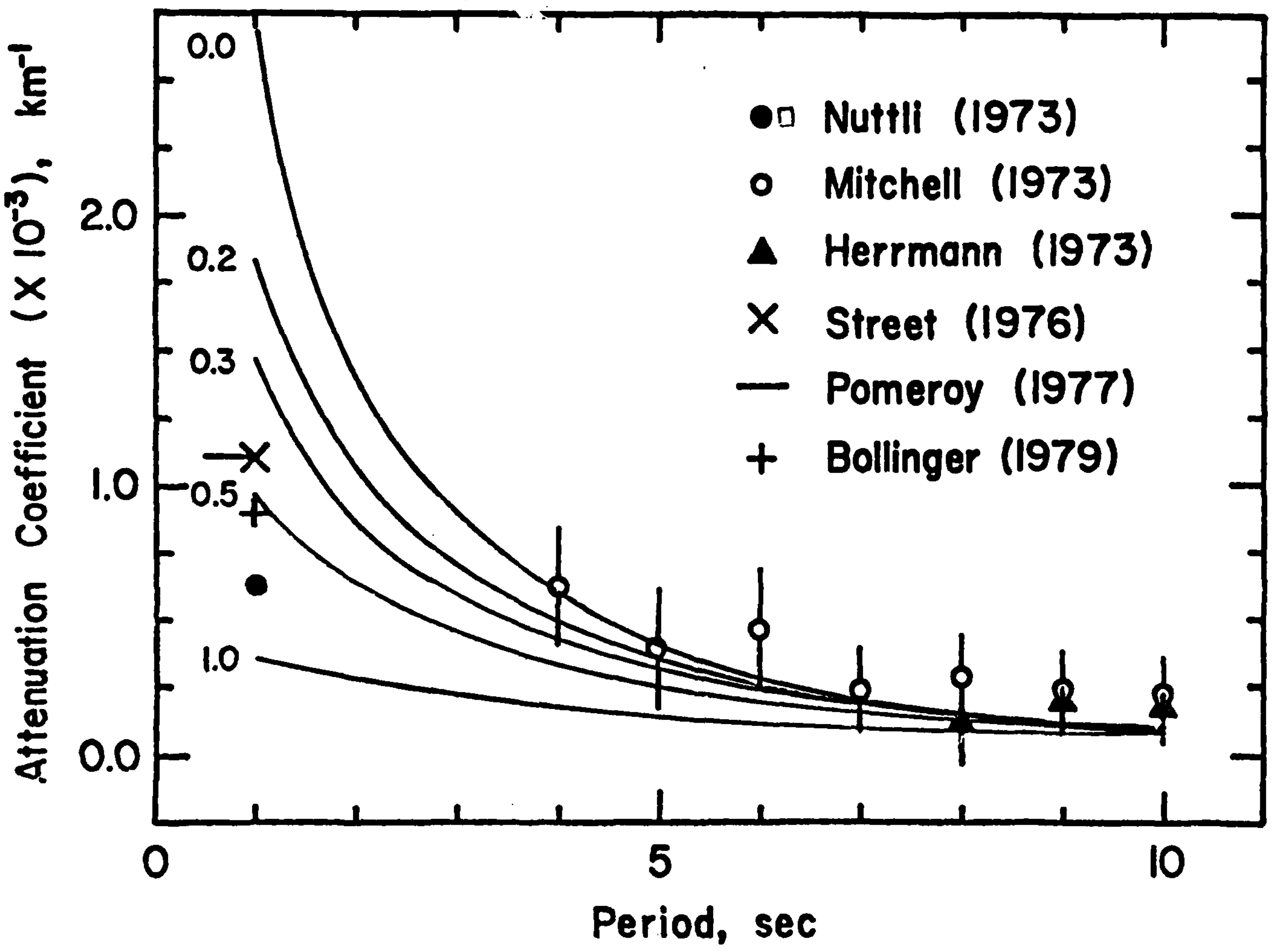


Figure 6. Higher-mode Rayleigh wave attenuation coefficient data and theoretical values obtained for various frequency-dependent internal friction models for which the exponent of frequency (ζ) is assumed to be uniform over the entire frequency range. From Mitchell (1980)

of the absence of data at periods between 1 and 4s. For example, the data could be explained equally well by models in which Q_{β}^{-1} increases uniformly with increasing period or in which Q_{β}^{-1} exhibits a maximum at periods of 2-3s (Aki, 1980). A Q_{β}^{-1} model in which ζ is constant over the entire period range and one with a peak in Q_{β}^{-1} at a period a 3s appear in Figure 7.

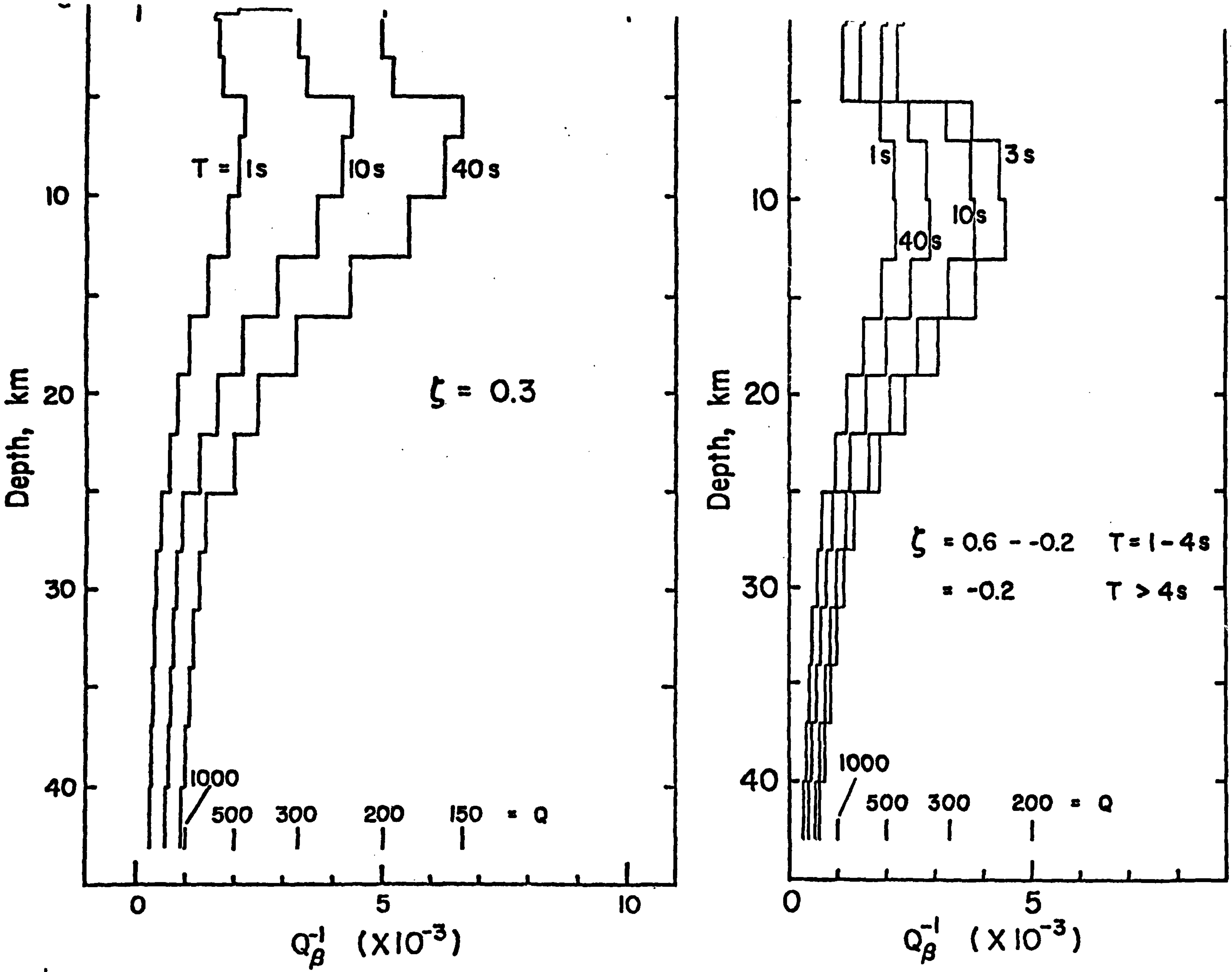


Figure 7. Two frequency-dependent Q_β^{-1} models of the crust in the eastern United States. From Mitchell (1980).

Regionalized Q_β^{-1} Models and Shear Wave Attenuation

Q_β^{-1} models of the crust and upper mantle to depths of 200 km or more are now available for several regions of the world. It is interesting to compare published measures of body wave attenuation, such as Q_{scs} and t*, to values calculated for models of Q_β^{-1} obtained from surface waves. If variations in the surface wave models would explain lateral variations reported for t* or Q_{scs}, then it would mean that significant lateral variations of Q_β are restricted to roughly the upper 200-300 km of the mantle.

In oceanic regions I have calculated values of Q_{scs} to be expected for the following regions of the Pacific: (1) >100 m.y. in age, (2) 50-100 m.y. in age, (3) the Nazca plate, and (4) the East Pacific Rise. If a constant value of 170 is assumed for Q_β at all depths in the mantle greater than 250 km, the calculated values of Q_{scs} are 176 for the region >100 m.y. in age, 171 for the region 50-100 m.y. in age, 148 for the Nazca plate, and 132 for the East Pacific Rise. These values can be compared to the observed Q_{scs} values of Sipkin and Jordan (1980) which are 155-180 for old oceans and 135-140 for young oceans. This satisfactory agreement implies that lateral variations of Q_β in the Pacific could be, but are not necessarily, confined to the upper 250 km of the crust and mantle.

In continental regions, it is most convenient to utilize the difference in the value of t* between the eastern and western United States. Using the models of Lee and Solomon (1975), that difference is about 2.8 with the more highly attenuating western United States having the higher value. Solomon and Toksoz (1970) determined t*

values across the United States finding large regional differences. The value over most of the western United States appears to exceed that of the eastern United States by 5 or more. Failure of the t* values calculated for surface wave models to explain the large observed variations suggests that regional variations in Q could extend to depths in the mantle much greater than 250 km. Differences in the models of Lee and Solomon (1975) between the eastern and western United States do indeed extend to much greater depths than 250 km.

References

Aki, K., Attenuation of shear waves in the lithosphere for frequencies from 0.05 to 25 Hz, Phys. Earth Planet. Int., in press, 1980.

Bollinger, G.A., Attenuation of the Lg phase and the determination of m_b in the Southeastern United States, Bull. Seism. Soc. Am., 69, 45-63, 1979.

Canas, J.A., Rayleigh Wave Propagation and Attenuation across the Atlantic Ocean, Ph.D. Thesis, Saint Louis University, 238 pp., 1980.

Canas, J.A., and B.J. Mitchell, Lateral variation of surface-wave anelastic attenuation across the Pacific, Bull. Seism. Soc. Am., 68, 1637-1650, 1978.

Canas, J.A., B.J. Mitchell, and A.M. Correig, Q_β^{-1} models for the East Pacific Rise and Nazca Plate, Proc. of the Advanced Seminar in Geophysics, University of Newcastle upon Tyne, in press, 1980.

Cheng, C.C., Determination of Crustal Q Structure from Multi-mode Surface Waves, Ph.D. Thesis, Saint Louis University, in preparation, 1980.

Cheng, C.C, and B.J. Mitchell, Q structures of the Colorado Plateau and the Basin and Range Province from higher-mode surface wave observations (abst.), 1978 Midwest AGU meeting, Saint Louis University, 1978.

Correig, A.M., and B.J. Mitchell, Regional variation of Rayleigh wave attenuation coefficients in the eastern Pacific, Pure and Appl. Geophys., in press, 1980.

Hasegawa, H.S., Surface- and body-wave spectra of Cannikin and shallow Alentian earthquakes, Bull. Seism. Soc. Am. 63, 1201-1225, 1973.

Herrmann, R.B., Surface wave generation by the south central Illinois earthquake of November 9, 1968, Bull. Seism. Soc. Am., 63, 2121-2134, 1973.

Herrmann, R.B., and B.J. Mitchell, Statistical analysis and interpretation of surface wave anelastic attenuation data for the stable interior of North America, Bull. Seism. Soc. Am., 65, 1115-1128, 1975.

Lee, W.B., and S.C. Solomon, Inversion schemes for surface wave attenuation and Q in the crust and the mantle, Geophys. J. Roy. Ast. Soc., 43, 47-71, 1975.

Lee, W.B., and S.C. Solomon, Simultaneous inversion of surface wave phase velocity and attenuation: Love waves in western North America, J. Geophys. Res., 83, 3389-3400, 1978.

Mitchell, B.J., Radiation and attenuation of Rayleigh waves from the southeastern Missouri earthquake of 21 October 1965, J. Geophys. Res., 78, 886-899, 1973a.

Mitchell, B.J., Surface wave attenuation and crustal anelasticity in central North America, Bull. Seism. Soc. Am., 63, 1057-1071, 1973b.

Mitchell, B.J., Regional Rayleigh wave attenuation in North America, J. Geophys. Res., 80, 4904-4916, 1975.

Mitchell, B.J., Anelasticity of the crust and upper mantle beneath the Pacific Ocean from the inversion of observed surface wave attenuation. Geophys. J. Roy. Ast. Soc., 46, 521-533, 1976.

Mitchell, B.J., Frequency dependence of shear wave internal friction in the continental crust of eastern North America, J. Geophys. Res., in press, 1980.

Mitchell, B.J., L.W.B. Leite, Y.K. Yu, and R.B. Herrmann, Attenuation of Love and Rayleigh waves across the Pacific at periods between 15 and 110 seconds, Bull. Seism. Soc. Am., 66, 1189-1201, 1976.

Mitchell, B.J., N.K. Yacoub, and A.M. Correig, A summary of seismic surface wave attenuation and its regional variation across continents and oceans, Geophys. Mono. 20, The Earth's Crust, edited by J.G. Heacock, 405-425, AGU, Washington, D.C., 1977.

Nuttli, O.W., Seismic wave attenuation and magnitude relations for eastern North America, J. Geophys. Res., 78, 876-885, 1975.

Pomeroy, P.W., Aspects of seismic wave propagation in eastern North America, A preliminary report, Roundout Associates, 1977.

Sipkin, S.A., and T.H. Jordan, Regional variation of Q_{ScS}, Seism. Soc. Am., in press, 1980.

Solomon, S.C., and M.N. Toksoz, Lateral variation of attenuation of P and S waves beneath the United States, Bull Seism. Soc. Am., 60, 819-838, 1970.

Street, R.L., Scaling northeastern United States/southeastern Canada earthquakes by their Lg waves, Bull. Seism. Soc. Am., 66, 1525-1537, 1976.

Yacoub, N.K., and B.J. Mitchell, Attenuation of Rayleigh wave amplitudes across Eurasia, Bull. Seism. Soc. Am., 67, 751-769, 1977.

12. EFFECTS OF ATTENUATION ON SURFACE WAVES

by

Thomas C. Bache

Systems, Science and Software
P. O. Box 1620
La Jolla, California 92038

Introduction

I was asked to write some comments under this topic. We have not done any substantial research at Systems, Science and Software (S^3) regarding the effects of attenuation on surface waves, but have been using data collected by others to compute synthetic seismograms. I will comment on that data we have found most useful and on the influence of attenuation on our seismograms. Our best description of our work is included in a paper by Bache, Rodi and Harkrider (1978) and a report by Bache, Rodi and Mason (1978). Results from these reports are discussed extensively under Topic 13.

We also refer to a summary report on the anelastic properties of the crust by Mitchell and Bache (1977). The main purpose of that report was to compare the anelastic properties of the western United States to those in the eastern United States and stable regions of Eurasia. No new results were presented, but results from many publications were summarized. Conclusions relevant for surface wave studies were that surface wave amplitudes at periods near 20 seconds are mainly influenced by the shear wave internal friction (Q_β^{-1}) values in the lower crust. These values are much lower and more uniform than the Q_β^{-1} values in the upper crust. Consequently, the amplitudes important for M_s exhibit relatively little variation due to anelastic attenuation from one continental region to another.

Q Models from $\gamma(\omega)$

Measurements of surface wave attenuation are generally in terms of $\gamma(\omega)$, where the attenuation is assumed to be exp $[-\gamma r]$. If the structure is known, the γ data can be inverted to determine a Q model. This is a well-known technique used, for example, by Mitchell (1975).

Under Topic 13 we summarize the results of a study of NTS explosions recorded at ALQ and TUC. Crustal models were determined for the NTS-ALQ and NTS-TUC paths in that study. We used the western United States attenuation data of Mitchell (1975) to infer Q_β models consistent with those crustal models. These Q_β models are shown in Figure 1 together with the associated attenuation for the relevant ranges. At these ranges attenuation is not very important at the long periods of primary interest.

Surface Wave Synthetic Seismograms and Conventional Q Models

The Bache, Rodi and Harkrider (1978) study summarized under Topic 13 used a method that virtually guaranteed that synthetic seismograms would have the same group velocity dispersion as the observations at ALQ and TUC. But the waveform agreement, shown in Figure 2, shows that more than the dispersion is in agreement. The amplification of the model, the source spectrum and the attenuation are also about right (though they could include cancelling errors). This is an indication that our conventional ideas about these factors are not far from reality.

Bache, Rodi and Mason (1978) studied the comparison of synthetic and observed amplitudes at ALQ and TUC. Again, the major results are summarized under Topic 13. An important result relevant for consideration of attenuation is that the source level inferred from the TUC observations is about 1.5

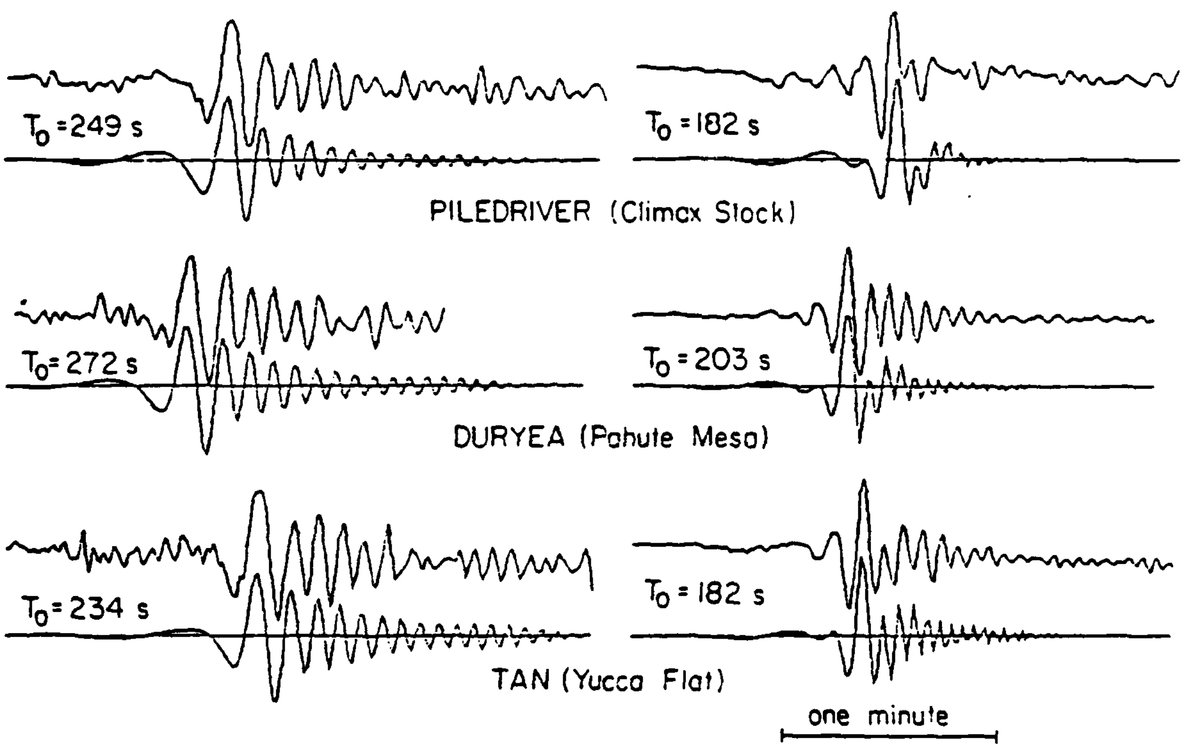


Figure 2. Theoretical and observed seismograms are compared at ALQ (left) and TUC for events in three test areas at NTS. A bar indicating one minute is shown. In each pair the observed (top) and theoretical records start at the same time with respect to the explosion detonation and this time is indicated as T_0.

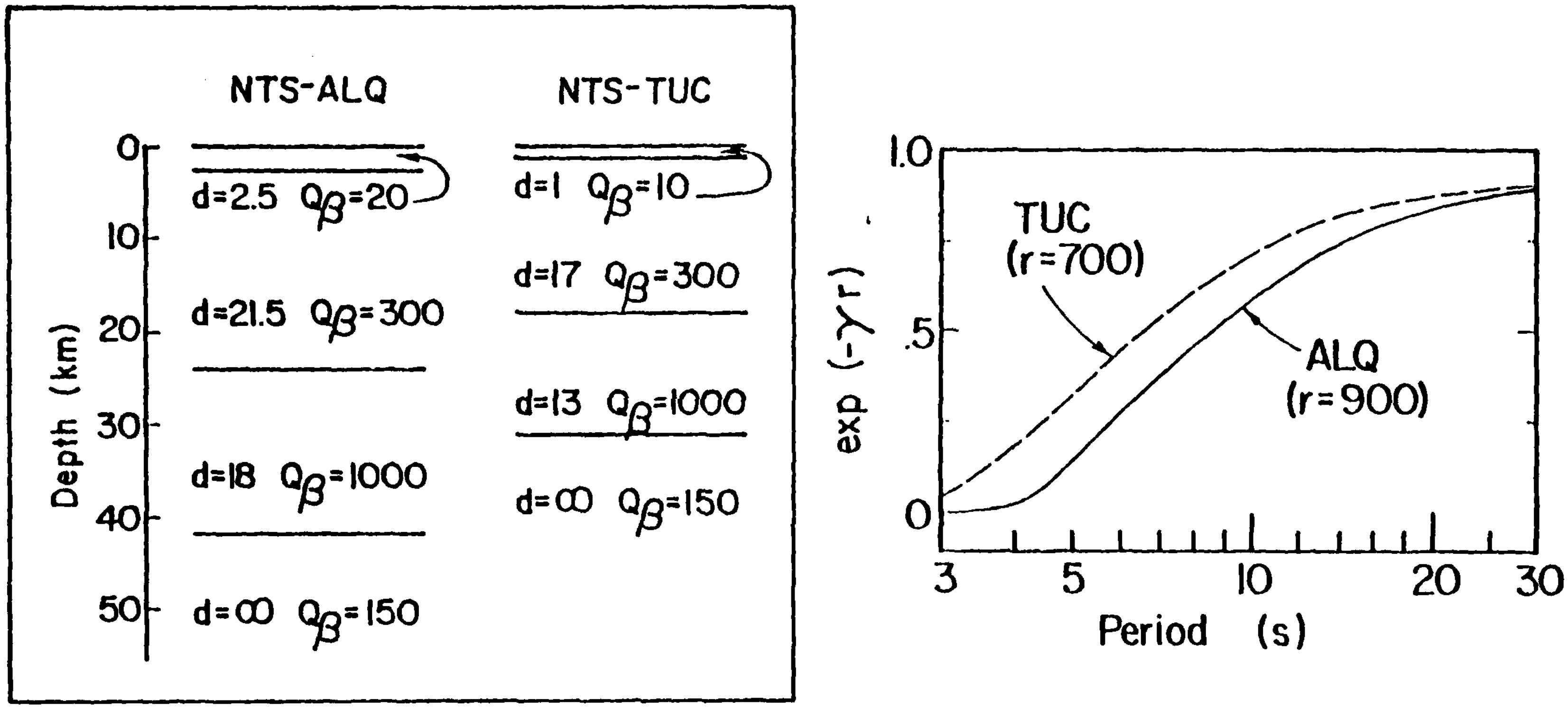


Figure 1. The Q_β models used to derive $\gamma(\omega)$ are shown at left for the NTS-ALQ and NTS-TUC paths with d denoting layer thickness. At the right the amplitude attenuation is plotted for the two paths.

times larger than that inferred from the ALQ observations. However, we conclude that only a portion of this discrepancy can plausibly be accounted for by assuming larger γ for NTS-ALQ and/or smaller γ for NTS-TUC. That is, if we assume that γ is in the range measured in studies like that of Mitchell (1975).

Conclusions

We have not done very much work with surface wave attenuation. We do find that available γ data lead to plausible Q models which can be used with models for the source and crust to compute synthetic seismograms that match the data. We have not been concerned with the frequency dependence of the crustal Q_β which influences surface waves. We also have not looked carefully at regional variations in Q, other than to conclude that they do not appear to be very great among continental structures.

There are not a great deal of available γ data. We have relied mainly on the work by Mitchell and his coworkers for the United States (Mitchell, 1973, 1975; Herrmann and Mitchell, 1975). We have not yet been synthesizing surface waves for Eurasian events, but plan to use results published by Burton (1977) and Yakoub and Mitchell (1977).

The entire matter of appropriate Q models is a subject for extensive research at this time. The state-of-the-art of understanding of Q seems to be rapidly improving.

References

Bache, T. C., W. L. Rodi and D. G. Harkrider (1978), "Crustal Structures Inferred from Rayleigh-Wave Signatures of NTS Explosions," BSSA, 68, pp. 1399-1413.

Bache, T. C., W. L. Rodi and B. F. Mason (1978), "Source Amplitudes of NTS Explosions Inferred from Rayleigh-Waves at Albuquerque and Tucson," Systems, Science and Software Report SSS-R-78-3690 submitted to ARPA/VSC, June, 91 pages.

Burton, P. M. (1974), "Estimations of Q_{γ}^{-1} from Seismic Rayleigh Waves," Geophys. J., 36, pp. 167-189.

Herrmann, R. B. and B. J. Mitchell (1975), "Statistical Analysis and Interpretation of Surface-Wave Anelastic Attenuation Data for the Stable Interior of North America," BSSA, 65, pp. 1115-1128.

Mitchell, B. J. (1973), "Surface-Wave Attenuation and Crustal Anelasticity in Central North America," BSSA, 63, pp. 1057-1071.

Mitchell, B. J. (1975), "Regional Rayleigh Wave Attenuation in North America," JGR, 80, pp. 4904-4916.

Mitchell, B. J. and T. C. Bache (1977), "Some Evidence Bearing on the Nature and Variability of the Anelastic Properties of the Upper Mantle Beneath North America and Eurasia," Systems, Science and Software Report SSS-R-77-3164 submitted to ARPA/VSC, March, 28 pages.

Yacoub, N. K. and B. J. Mitchell (1977), "Attenuation of Rayleigh-Wave Amplitudes Across Eurasia," BSSA, 67, pp. 751-769.

End 6-82